



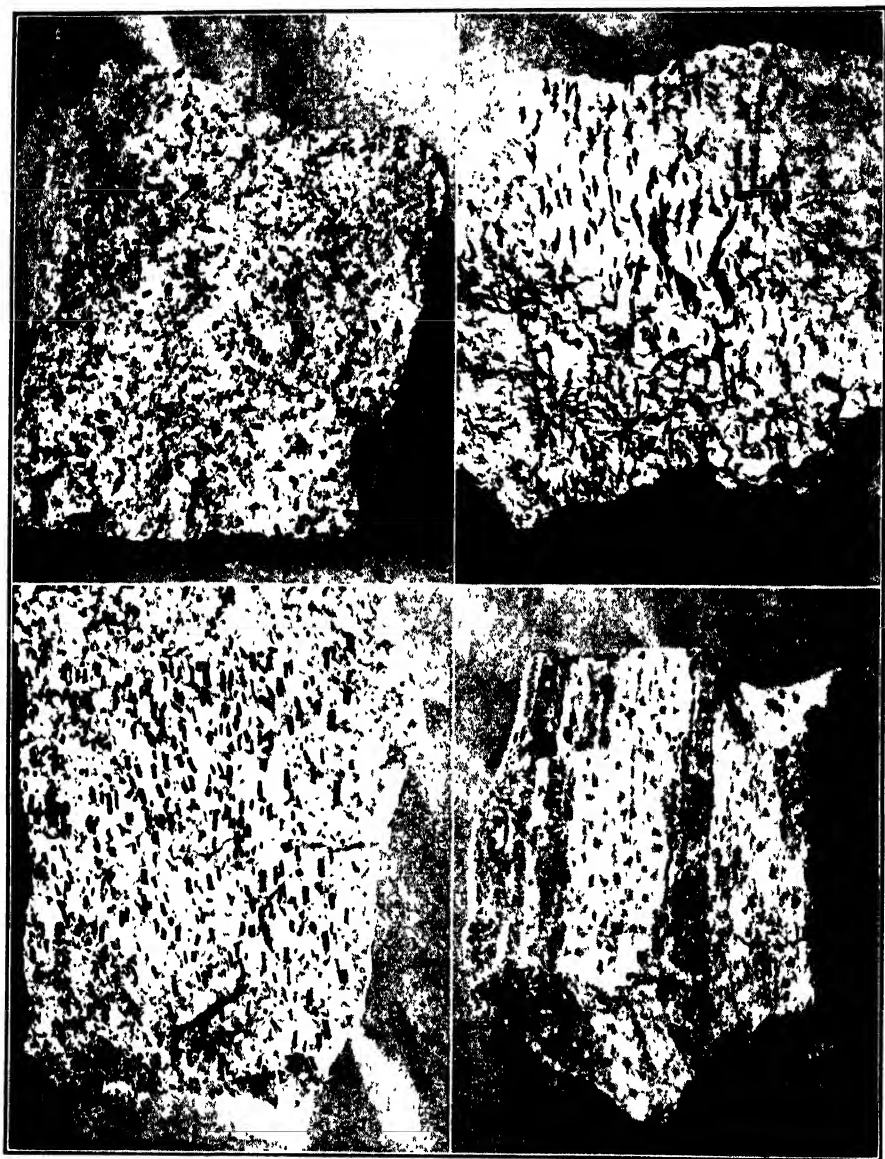
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PETROGRAPHY AND PETROLOGY

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(Frontispiece)

Four phases of syenite, Snowbank Lake, Minnesota. About one-half natural size. An example of petrographic interpretation. Upper left, syenite, massive; upper right, syenite gneiss (metamorphic); lower left, gneissoid syenite or trachitoid syenite (primary gneiss); lower right, syenite injection gneiss, lit-par-lit. There may be some question whether a gneiss like that shown in the upper right-hand illustration may be produced by piezocrystallization of magma as well as by metamorphism.

PETROGRAPHY AND PETROLOGY

A Textbook

BY

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*The quality of the materials used in the manufacture
of this book is governed by continued postwar shortages.*

To
CHARLES PETER BERKEY
INSPIRING TEACHER OF THE
NEW PETROLOGY

PREFACE

It is the purpose of this textbook to combine a moderately advanced laboratory study of rocks with a classroom discussion of the origin of the features seen in the laboratory. The two supplement each other. It should hardly be necessary to add that the laboratory work should be based on good illustrative material—rocks and thin sections. No textbook, however perfect, is a substitute for direct observation.

The general tone of the book is perhaps conservative or even reactionary in the matter of rock classifications, for there have in recent years been several good suggestions of progressive changes and reforms. The attempt here is to be practical and follow present-day usage.

The descriptive and systematic side of petrography is now so far advanced that few essential features of igneous rocks have escaped the careful petrographers. As a result, descriptive petrography is giving way to interpretative petrology. Reading the life history of a rock is a much higher achievement than a description and classification. In this work, therefore, attention is directed first and rather briefly to the features to be expected in rocks, and then to some of the theories by which these features are interpreted in terms of history. This interpretative branch of the subject, moreover, is treated here only in a preliminary way as applied to common rocks, and not at all adequately for the altered rocks associated with ore deposits.

The discussions of igneous rocks are longer than those of sediments and metamorphic rocks. If special theories as to the inside of the earth and its processes apply to the discussions of both igneous rocks and others, the early statements are referred to rather than repeated.

There is now so much literature describing the occurrences and varieties of rocks that any summary of it is too voluminous for use in a textbook. The student may find these summaries in more encyclopedic works.

Few references are given in the main body of the text, but some standard books are listed on page xi, and a series of selected readings are listed at the end of the book for the use of those who wish to go farther in the study of certain topics. The references and illustrations are chiefly those of easy access. Where the list is long, a selection from the first part of the list may be sufficient. Readings and leading questions may well be assigned in a course while the lectures and laboratory work are devoted to the description of the several rock families. Later the petrologic theories may be discussed while the laboratory work proceeds to the practical applications of petrography.

For petrology and even for petrography the student needs a theory as to processes not seen. His success in interpretation and even in description depends on his having the most reasonable theory. It is noteworthy that scientific theories are developed by two rather distinct methods: deduction, that is, assuming a hypothesis and inferring what the consequences must be; induction, that is, finding a body of facts and building a hypothesis from them. Modern science uses both with more or less system. The process is: (1) Find the facts, (2) classify them, (3) turn loose the imagination to build hypotheses, (4) reason out the consequences of each, and (5) test the hypotheses by finding whether the reasoned consequences are facts. The system is thus cyclic and yields constantly more facts and improved theories.

In the parts of this work devoted to petrology there are given some of the basal facts and several prominent theories as to the origin of rocks and the causes of their diversity of composition and form. It is of the utmost importance that observation and description should be clearly distinguishable from theories in any scientific text. The student is thus encouraged to judge for himself the relative values of suggestions and is free to draw his own conclusions from the facts presented. This separation of fact and theory is of course difficult, especially if the theory is widely accepted and a terminology has been built up around it. It is hoped that such terms implying a knowledge of inferred processes have not been used too freely throughout the book.

FRANK F. GROUT.

MINNEAPOLIS, MINN.,
April, 1932.

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Foremost among the teachers to whom the author owes much is Dr. Charles P. Berkey. Dr. Berkey's teachings may not be accurately reflected in the statements of this book, but his practical common-sense attitude has had a great influence upon the author's general philosophy of the subject as well as upon the details of petrographic usage and nomenclature. The influence of Prof. L. V. Pirsson and that of Prof. Joseph Barrell will also be apparent. Many published works have been drawn upon freely.

The author is glad also to acknowledge his great indebtedness to his colleagues at the University of Minnesota. Dr. W. H. Emmons, Dr. G. A. Thiel, Dr. J. W. Gruner and Dr. G. M. Schwartz have given the benefit of general discussions and contributed certain special compilations. Drawings have been made by Mr. H. R. Norman, Mr. W. I. Gardner, Miss Faith Patterson and Miss Virginia Dustin. Several of the photographs were made during field work of the author on the Minnesota Geological Survey. Others were made by Mr. Henry Morris, and still others are credited where they appear.

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PETROGRAPHY AND PETROLOGY

PART I

INTRODUCTION

A rock is anything that makes up one of the structural units of the earth's crust; the term is also used for a specimen representative of such a unit. It must have sufficiently definite composition and character so that it can be defined as a unit and distinguished from other units, though its contacts may be gradational, and its limits arbitrarily set. In size rocks may differ greatly, and no one with wide experience in dealing with rocks is inclined to set a lower limit and say that smaller units are not properly rocks. In the popular mind a rock is a hard solid, but from a scientific point of view beds of loose sand and volcanic ash must be included as units of the earth's crust as truly as a granite or a chert.

Petrography is here regarded as the systematic and descriptive side of the study of rocks, whereas *petrology* is more comprehensive, covering not only the description but also the theories of origin of rocks—petrogeny—and the interpretation of the facts in petrography. It deals with rocks as the documents of geologic history.

Both terms are more commonly used for microscopic work than for work on hand specimens—rock study. Microscopic work on rocks was also formerly distinguished as lithology. Petrography and petrology are interwoven with and are parts of the broader science of geology. Descriptive petrography rests largely on a foundation of mineralogy, but interpretative petrology finds structures and textures fully as significant as minerals.

NATURE OF CRITERIA

At various points in the book are lists of criteria by which certain standard rocks and processes are recognized. The main criteria in rocks are structures, textures, and minerals in various combinations. These are strictly limited in number, but their various combinations are almost infinite. Certain combinations are very definite indications that a particular process must have taken place in the history of the rock. More often the combinations found leave one in doubt as to which of several processes may have occurred. The criteria may suggest a certain

event or sequence of events, but leave the possibility open, that some other events may have given the result. The criteria listed in this work include many that are of this suggestive type and that are not in themselves conclusive.

The use of a great variety of criteria by men of long experience is almost instinctive; they hardly know what points they have observed in coming to their conclusions. Just as a miner recognizes galena "at sight," so the older petrographers know the process that has affected rocks at sight but find it hard to explain why or how they know. The beginner is not able to do this without a table of criteria, just as a beginner in mineralogy uses a table to identify galena.

The importance of listing such indefinite criteria is found in their cumulative value. If a whole series of suggestive criteria point the same way, the conclusion based on them may be just as certain as if based on one absolute criterion. This may perhaps be illustrated by the signs of rain.

CRITERIA SUGGESTING RAINY WEATHER

A. Falling barometer:

- Rheumatic aches.
- Smell of a storm (liberated marsh gas).

B. High humidity:

- Ring around the moon; size of it.
- Sun drawing water (visible rays).
- Cloudiness, especially if increasing and especially if developed in midday.
- Leaves show their under sides.
- Hair stretches; curly hair kinks.
- Certain dyes change color.
- Doors and drawers stick.
- Salt is damp; tobacco is wet.
- Red sky in the east in the morning.
- Stars "huddle," twinkle, hide, or get hazy.

C. Animals are restless:

- Robins whistle for rain.
- Flies bite extra hard and flock into houses.
- Chickens oil their feathers.
- Sea birds stay on land.
- Cats and dogs nibble grass.
- Ants are especially busy.
- Spiders come out.
- Cats lick their fur the wrong way.

D. Weather prediction by the Weather Bureau:

- Based largely on the approach of a "low."

E. Miscellaneous:

- East wind, especially if long continued.
- Clear morning after night's rain.
- Mackerel sky.
- Fog in the morning goes up.
- Cloudy in the west at evening.
- Sun shower the day before.
- Rainbow in the morning.
- Smoke drifts low along the ground.
- Thunder, especially if growing louder and nearer.
- Visible shower approaching.

Someone holding up an umbrella in the distance.

Many stars visible.

If one sees or hears distant things clearly.

Most persons forecast the coming weather by accumulating half a dozen or so of these signs or criteria, all pointing the same way.

For most common rocks there is no tabulation of criteria as such. Instead, the criteria are given as formulas and definitions in the descriptive text. For certain special cases tabulated criteria have seemed useful in distinguishing rocks or special features in the rocks. For the most part the tables of criteria refer to processes that have acted upon the rock at some time in the history of its development.

GENERAL GEOLOGY OF ROCKS

The subject of general geology should normally precede a careful study of petrography and petrology. Even to those who have not had this general training, however, the formation of rocks is a matter of everyday observation. The wind blows sand into dunes, and the rain washes mud into lakes or into rivers, which carry it toward the sea. The waves and currents of the sea sort and deposit the sands. The ocean and the land support their myriads of living beings, which deposit immense quantities of organic remains. Lake and spring waters by action of the air develop precipitates. Glaciers scour out great valleys and deposit moraines of the débris. Cliffs break down to talus slopes, forming volumes of rubble and breccia. Volcanoes pour out lava and scatter the fragments from their explosions. All these visible and familiar activities of geologic agents result in the formation of rocks (Figs. 113 and 202).

Nevertheless, when we study in detail the rocks produced by these visible processes, we find that they do not represent all the kinds of rocks found in the crust of the earth. The rocks that differ from them probably originated deep in the earth, where no human eye could watch their development. The details of the process in particular cases can be estimated by long and careful study of the history of the associated formations and by laboratory work producing artificial rocks; but it should be clearly understood that the processes are inferred and not observed. A large portion of the rocks now visible at or near the surface of the earth attained their present general nature at great depths and came to light only after a long geologic time of uplift and erosion. The proof that many igneous masses have once been liquid is found in the manner of their occurrence, penetrating fractures and enclosing blocks of their walls.

A detailed statistical study of the rocks exposed at the surface indicates that much of the bedrock nearest the surface has originated by transportation in water and deposition from suspension or from solution. Almost three-fourths of the rock exposures at the surface are such sediments. The rocks formed by the action of water, therefore, deserve a prominent place in any general petrography.

A different point of view, however, is obtained from a study of the structure of the earth's crust—the "crust" being defined as an outer layer of the earth approximately 10 miles thick. This crust is found to have in most places only a thin layer of sediments, if any, and the underlying rocks are largely those formed by the cooling of a hot liquid, the "magma," such as the deep-seated liquid supplying the lava for volcanoes. It is estimated that the proportions of rock in the crust are very close to the following:

	Per Cent
From magma	95
From sedimentary mud	4
From sedimentary sand.	0.75
From the calcareous hard parts of organisms.....	0.25

Some rocks of each of these groups may be altered so that they are best classified as metamorphic. All the other rocks together constitute an insignificant fraction of the whole crust, and so the dominance of the first group justifies the emphasis usually placed upon it. The magmatic products should logically be taken up first, also, because other rocks are largely derived from them, and because many ore deposits are related to the magmas from which these rocks form.

CLASSIFICATION OF ROCKS

Development of Schemes.—The early history of petrography as a science is largely a record of its differentiation from mineralogy and general geology. In the middle of the eighteenth century Linnæus divided the mineral kingdom into rocks, minerals and fossils. It was a hundred years later, in 1862, that Von Cotta in the second edition of his *Gesteinslehre* first proposed the main groups of rocks based on origin, in very much the sort of terms that are now used.

- I. Eruptive rocks.
- II. Sedimentary rocks.
- III. Metamorphic rocks.

Chemical and mineralogic methods of study were the foundations of nearly all work up to that date. Then came the era of the microscope, which has dominated the study of rocks ever since, with many important developments, of course, in the chemical and mineralogic sides of the field (see readings suggested). The advance of the systematic study of rocks since 1860 has been rapid, and many important suggestions need attention.

The Rosenbusch type of classification may be taken as an example of the type that dominated the petrographic world for many years near the close of the nineteenth century. Rosenbusch, studying chiefly igneous rocks, first classified them on the basis of structure, later on genetic characters, both in combination with mineral composition.

Any scheme of classification should be governed by certain general principles. In the first place no natural absolute division lines are to be found. Everywhere there are gradations. Igneous rocks even grade into aqueous rocks and may be weathered until hardly distinguishable from sediments. Both of these grade into metamorphic rocks by indistinguishable stages. Division lines must be arbitrary in any system.

In the second place the purpose of the classification should be kept in mind. It may well happen that two or more different schemes of classification should be employed to serve different purposes. One worker may describe a rock for the sake of a field name, another for a detailed microscopic classification, another for a study of its magma chemically, another to interpret its history, and still another only as a foundation on which to settle some economic problem.

The Point of View in Petrography.—During the development of scientific schemes of classification, petrographers have commonly been enthusiastic in their search for fresh, clear rock specimens—those that would serve as type examples, filling new pigeonholes in the scheme. Many of these are so rare as to be of little interest except to the systematist.

On the contrary, the geologist who tries to apply his petrography to economic problems *rarely finds fresh rocks*. It is almost axiomatic that the rocks associated with ore deposits are altered, and that the extent and nature of the alteration are much more important for the petrographer than the detailed systematic classification. He studies fresh typical rocks first, for a brief time, so as to be able to interpret the alteration and history of the rocks that he has to work with later.

Again, a contrast may be drawn between the field geologist, who classifies rock bodies emphasizing their occurrence and relations, and the laboratory man, who classifies hand specimens. The Rosenbusch classification of igneous rocks is an example of a scheme requiring field knowledge of the relations of a rock before classification, though in actual practice all who have had much experience in its use know that the determination of the relations of a rock is commonly left to the unreliable supposition that texture indicates occurrence. This practice of classifying a rock by one scheme when field information is available and by another scheme, using the same names, when only laboratory data are available is not to be recommended. The truth is that most petrographers who name rocks according to the geologic or field basis are using, unconsciously perhaps, two or more classifications.

It is clear that three systems of classification are justifiable, namely, geologic, petrographic, and petrologic. Each is entitled to full recognition; difficulty arises only in attempts to mix them. It is here assumed that the several classes based on geologic age and forms of occurrence are familiar to the student. The main emphasis will here be placed on a

strictly petrographic classification, based on the kinds and amounts of minerals, and the texture—characters which define in a material way the rocks under consideration. Some discussion will later be given to the petrologic features of rocks and the names based on genetic relations.

The Three Main Classes of Rocks.—The accepted division of rocks into three main classes is based on origin.¹

I. Igneous rocks are those formed from cooling magma. Magma is a natural liquid in or on the earth, most of it very hot, made up largely of a mutual solution of silicates, but with some oxides and sulphides, etc., and nearly always containing water and other gases held in solution by pressure (see pages 143–145).

The original term eruptive might be better than igneous, considering its derivation, but the term is now restricted to flows in contrast with intrusive rocks which are called irruptive. Pyrogenic is also used as equivalent to igneous.

In one sense ice, especially in polar regions, is an igneous rock, and water is a magma, but its temperature of crystallization is so far removed from the high temperatures characteristic of most igneous action that the bare mention of it suffices.

In a certain sense, also, the liquid magma and water may be considered rocks, but this extension of the term is questionable, because, although they occur in or on the earth's crust, their fluidity is opposed to our notions of the characteristics of a crust. Many of the accepted rocks of the earth, however, are soft or noncoherent.

II. Sedimentary rocks are formed through the agency of water, wind, or glacial ice and are deposited on the surface or in the outer crust of the earth at normal temperatures. The material may be detrital or precipitated. Far the largest part of the sediments is deposited by water in the sea, but a smaller part is deposited by water on or under the surface of the land. A still smaller fraction is deposited by wind and ice, forming eolian and glacial rocks. All are classed as mechanical deposits except a small percentage precipitated from solution in water. These precipitates are not sediments in the sense of mechanical deposits, but are included here because they are aqueous and are intimately associated with other sediments.

III. Metamorphic rocks are those that have been greatly changed in minerals or texture or both since their first formation as rocks. The changes are largely due to pressure in the depths of the earth or to the intrusion of igneous magmas near by. These effects are not uniform, and the rocks produced grade into those from which they were formed without any sharp line of division.

In the practical work of petrography the first question that arises is how to determine whether a rock is igneous, aqueous or of some other origin. Few text books or discussions give any criteria for such a determination, but in a great many cases the determination is not easy. Most

¹ Crook has suggested a genetic grouping according to processes: (1) *endogenetic* for deep-seated processes—igneous, exudation and metamorphic rocks—(2) *exogenetic* for weathered, detrital, solution, and organic rocks.

petrographic discussion starts with an assumption that field occurrence or some other work has already settled this primary problem. On the contrary most petrographers can cite many experiences in which a careful survey of all available criteria was needed to arrive at a safe primary classification. After detailed study of the several main families of rocks in each of the three great classes, the *most common members are readily distinguished* by relatively simple criteria. Many specimens of the rarer sorts need the cumulative evidence of a number of complex criteria. The indefinite nature of some of these criteria is noted on page 2. It should be noted also that the criteria of igneous rock are not exactly the same as the criteria that the rock formed from a liquid magma of the same composition.

CRITERIA FOR DISTINGUISHING IGNEOUS, SEDIMENTARY AND METAMORPHIC ROCKS¹

Igneous. (See also the criteria of magmatic segregation, and others.)

1. *Interlocking grains*: mostly hard rocks.
2. Characteristic mineral composition. Nephelite, olivine, chromite, arfvedsonite, leucite, perthite, tridymite, etc.
3. *Massive*: irregular shape, columnar joints, spheroidal weathering, only rarely banded.
4. Textures: glassy, porphyritic, amygdaloidal, cellular, orbicular, trachitic, trachitoid, diabasic, graphic, coronas, miarolitic cavities, granitoid.
5. High in feldspar.
6. One kind of plagioclase not two unless zoned.
7. Enclosing rocks baked.
8. Gradation laterally into igneous rocks definitely identifiable.
9. Evenly distributed euhedral accessories.
10. Formation transgresses schistosity but seems influenced by it.
11. No fossils or strata.
12. Quartz crystals stubby.
13. Normal order of crystallization.
14. Chilled borders.

Sedimentary. (See also criteria of mechanical sedimentation.)

1. *Stratification* and sorting; mostly soft rocks.
2. *Textures*: *Fragmental*, *bolitic*, *pisolitic*, porous, reniform, botryoidal, rounded and sorted grains.
3. Wide lateral extent.
4. The common rocks, clay, sandstone, limestone, and coal.
5. Intercalations with known sediments and gradations to them laterally.
6. Fossils.
7. Quartz chiefly, or calcite chiefly, or high in Al_2O_3 minerals, if soft.
8. Formation shows marked and regular differences in composition.
9. Formation persistently follows folds.
10. Minerals of chemical and organic precipitates. Salts, gypsum, glauconite, vein minerals, coal, chert, and carbonates.
11. Textural changes in bands laterally and vertically.
12. Accessories: (a) absent, (b) rounded if present, (c) segregated in placers.
13. Several plagioclases.
14. Original textures (preserved in axes of folds): cross bedding, mud cracks, ripple marks, foot prints, etc.
15. Not usually extremely basic.
16. Alpha quartz is usually from water solution.
17. Concretions, septaria, stalactites, crusts, etc.

¹ A detailed tabulation of minerals to be expected in various occurrences is given by Wherry and Gordon, Philadelphia, Acad. Nat. Sci. Proc., August, 1915, pp. 426-457.

Metamorphic. (See also criteria of dynamic metamorphism.)

1. *Parallelism of mineral grains, interlocking grains, hard rocks.*
2. *Secondary cleavage independent of bedding.*
3. *Banding, foliation, schistosity.*
4. *Characteristic minerals, staurolite, andalusite, garnet, wollastonite, grossularite, tremolite, kyanite, sillimanite, cordierite, zoisite, talc, etc.*
5. *Distorted pebbles, fossils, crystals, etc.*
6. *Granulation, augen, mortar and flaser structures, hornfels textures.*
7. *Wavy extinction especially if very pronounced; distinguish vein quartz.*
8. *Bent cleavage, folds, faults.*
9. *Old look.*
10. *Presence of graphite in quantity (say 5 per cent).*
11. *Minerals as metacrysts (list, page 364).*
12. *Secondary enlargement, as quartz in quartzite.*
13. *Geographic and geologic position of outcrop: (a) in deeply eroded known pre-Cambrian regions, (b) next to rocks that may have affected it, (c) grading laterally into a recognizable metamorphic rock.*
14. *The crystalloblastic series.*
15. *High ratio of length to the thickness of biotite, chlorite and hornblende.*
16. *Rocks chiefly mica, chiefly chlorite, or hard rocks chiefly quartz or calcite*

PETROGRAPHIC METHODS AND MINERALS

The tendency of several decades to make petrography an elaboration of microscopic optical mineralogy is now somewhat outgrown, for geologists find that all methods and all branches of science may be applied to the study of rocks. Petrographic field methods are outlined for each of the three main classes of rocks later, as each group is described. Mineralogic methods include chemical, blowpipe, physical, and especially optical tests. Special petrographic methods include those for the separation of minerals and measurement of thin sections and powders, also some calculations based on chemical analyses (see pages 19 to 21 and list of further readings).

PREREQUISITE METHODS

Assuming that the methods of mineralogy and optical mineralogy are known, the petrographer applies these and other methods to the study of rocks. The methods most applied, and in fact prerequisite to this work, are in general as follows:

Mineralogic and crystallographic.

Hardness, color, luster, streak, fusibility, specific gravity, crystal form, cleavage, solubility, flame and bead reactions, magnetism, charcoal and tube tests.

Chemical.

Qualitative chemical tests, notably silica jelly and effervescence.
Quantitative chemical analyses.

Optical.

Use and adjustment of a petrographic microscope and accessories.
Making a thin section; and preparing powders for work with oils.
Cleavage angles.
Relief and shagreen.
Becke method for comparative indices.
Inclined illumination for comparative indices.
Production and use of polarized light.
Estimation of three indices, α , β , and γ .
Estimation of birefringence.

Angle between cleavage or crystal faces and vibration directions of α , β , and γ .

Color of each ray α , β , and γ .

Optic sign and sign of elongation.

Uniaxial or biaxial character.

Optic angle.

Dispersion.

Microchemical methods and stains: (1) on smooth surface, (2) on powder, (3) on streak.

Mineralographic study of polished surface.

The optical methods to be adopted vary somewhat with the material available and with the objects to be attained. A thin section requires different methods from a powder taken from selected parts of a hand specimen. In general the main groups of minerals and the texture are best studied in thin section, whereas accurate studies of minerals are made on the powder in a series of refractive oils; or if opaque the minerals may be studied in polished surfaces, with the blowpipe, or by the streak microscopically. The tables on pages 443 to 478 will often serve for the identification of common rock minerals. More elaborate tables should be consulted for detailed work.

COMMON MINERALS

The common minerals in the earth's crust fall into relatively few groups—10 may be selected to cover 99 per cent of all igneous material.

Quartz	Pyroxenes	Olivine	Magnetite
Feldspars	Micas	Nephelite	Apatite
Amphiboles		Leucite	

In a study of the whole crust others are almost negligible, but there are certain very widespread minor accessories and alteration products, and near the surface of the earth weathering has produced a local abundance of several other minerals, all of which should be very familiar to the student.¹

Iron oxides	Chlorite	Tourmaline	Spinel
Clay minerals	Epidote	Staurolite	Corundum
Carbonates	(Obsidian)	Cordierite	Perovskite
Chalcedony	Serpentine	Kyanite	Zeolite
Sulphides	Garnet	(Contact minerals)	Gypsum
Titanite	Talc	Rutile	Zoisite
Zircon			

The following table is intended to furnish a basis for certain calculations commonly made by petrographers. The figures first after the mineral name are the specific gravities. Next is a formula if a simple one can be written, and the percentages of each constituent as determined from the formula or by making an average of many analyses (see Problem 8, page 18).

¹ Iddings, J. P., "Rock Minerals," John Wiley & Sons, Inc., 1911. A. JOHANNSEN, "Essentials for the Determination of Rock-forming Minerals and Rocks," The University of Chicago Press, 1922. E. S. LARSEN, The non-opaque minerals, U. S. Geol. Survey Bull. 679, 1921. A. N. WINCHELL, "Elements of Optical Mineralogy," Part II, John Wiley & Sons, Inc., 1927.

The last two works include many rare minerals as well as those of common rocks.

TABLE I.—COMPOSITION OF ROCK MINERALS

(Modified from Hance)

Actinolite: 3.1—(OH) ₂ Ca(Mg, Fe) ₂ (Si ₄ O ₁₁) ₂ : CaO—13.0; MgO—15.0; FeO—12; SiO ₂ —58.0; H ₂ O—2.0.
Aegirite (acmite): 3.55—NaFeSi ₂ O ₆ : Na ₂ O—13.4; Fe ₂ O ₃ —34.5; SiO ₂ —52.1.
Albite: 2.63—NaAlSi ₃ O ₈ : Na ₂ O—11.8; Al ₂ O ₃ —19.4; SiO ₂ —68.8.
Almandine (garnet): 4.05—Fe ₃ Al ₂ Si ₂ O ₁₂ : FeO—43.2; Al ₂ O ₃ —20.5; SiO ₂ —36.3.
Alunite: 2.66—KAl ₃ O ₄ H ₂ S ₂ O ₅ : K ₂ O—11.4; Al ₂ O ₃ —37.0; SO ₃ —38.6; H ₂ O—13.0.
Analcite: 2.25—NaAlSi ₂ O ₆ H ₂ O: Na ₂ O—14.0; Al ₂ O ₃ —23.2; SiO ₂ —54.6; H ₂ O—8.2.
Andalusite: 3.18—AlSiO ₃ : Al ₂ O ₃ —62.9; SiO ₂ —37.1.
Andesine: 2.68—Ab ₂ An ₂ : Na ₂ O—6.9; CaO—8.3; Al ₂ O ₃ —26.6; SiO ₂ —58.2.
Andradite (garnet): 3.85—Ca ₃ Fe ₂ Si ₂ O ₁₂ : CaO—33.0; Fe ₂ O ₃ —31.5; SiO ₂ —35.5.
Anhydrite: 2.94—CaSO ₄ : CaO—41.2; SO ₃ —58.8.
Anorthite: 2.75—CaAl ₂ Si ₂ O ₈ : CaO—20.1; Al ₂ O ₃ —36.7; SiO ₂ —43.2.
Apatite: 3.20—Ca ₄ (Ca, F)P ₃ O ₁₂ : CaO—55.5; P ₂ O ₅ —42.3; F—3.8.
Augite: 3.31—complex: CaO—20.9; MgO—12.6; FeO—6.9; Fe ₂ O ₃ —4.6; Al ₂ O ₃ —7.4; SiO ₂ —47.6.
Barite: 4.45—BaSO ₄ : BaO—65.7; SO ₃ —34.3.
Bauxite: 2.48—Al ₂ O ₃ ·2H ₂ O: Al ₂ O ₃ —73.9; H ₂ O—26.1.
Biotite: 2.95—complex: K ₂ O—11.2; (Mg, Fe)O—19.2; Al ₂ O ₃ —24.4; SiO ₂ —43.1; H ₂ O—2.1.
Calcite: 2.72—CaCO ₃ : CaO—56; CO ₂ —44.
Chalcedony: 2.61—SiO ₂ ; SiO ₂ —100.
Chlorite (penninite): 2.72—complex: MgO—19.5; FeO—23.2; Al ₂ O ₃ —16.5; SiO ₂ —29.2; H ₂ O—11.6.
Cordierite: 2.63—H ₂ (Mg, Fe) ₄ Al ₂ Si ₁₀ O ₃₇ : SiO ₂ —49.4; Al ₂ O ₃ —33.6; FeO—5.3; MgO—10.2; H ₂ O—1.5.
Corundum: 4.0—Al ₂ O ₃ : Al ₂ O ₃ —100.
Diopside: 3.33—CaMgSi ₂ O ₆ : CaO—25.8; MgO—18.6; SiO ₂ —55.6.
Dolomite: 2.84—CaMgC ₂ O ₆ : CaO—30.4; MgO—21.7; CO ₂ —47.9.
Enstatite: 3.17—MgSiO ₃ : MgO—40; SiO ₂ —60.
Epidote (pistacite): 3.38—Ca ₂ (Al, Fe) ₂ HSi ₂ O ₁₂ : CaO—23.5; Al ₂ O ₃ —24.1; Fe ₂ O ₃ —12.6; SiO ₂ —37.9; H ₂ O—1.9.
Fayalite (olivine): 4.14—Fe ₂ SiO ₄ : FeO—70.6; SiO ₂ —29.4.
Fluorite: 3.13—CaF ₂ : Ca—51.1; F—48.9.
Forsterite (olivine): 3.25—Mg ₂ SiO ₄ : MgO—57.1; SiO ₂ —42.9.
Gibbsite: 2.36—Al(OH) ₃ : Al ₂ O ₃ —65.4; H ₂ O—34.6.
Grossularite (garnet): 3.57—Ca ₃ Al ₂ Si ₂ O ₁₂ : CaO—37.3; Al ₂ O ₃ —22.7; SiO ₂ —40.0.
Gypsum: 2.32—CaSO ₄ ·2H ₂ O: CaO—32.6; SO ₃ —46.6; H ₂ O—20.9.
Hedenbergite: 3.6—CaFeSi ₂ O ₆ : CaO—22.2; FeO—29.4; SiO ₂ —48.4.
Hematite: 5.1—Fe ₂ O ₃ : Fe—69.9; O—30.1.
Hornblende: 3.24—complex: CaO—5.8; MgO—8.3; FeO—22.1; Fe ₂ O ₃ —16.3; Al ₂ O ₃ —10.5; SiO ₂ —37.0.
Hypersthene: 3.45—(Mg, Fe)SiO ₃ : MgO—17.3; FeO—30.9; SiO ₂ —51.8.
Kaolinite: 2.62—Al ₂ Si ₂ O ₇ ·2H ₂ O: Al ₂ O ₃ —39.5; SiO ₂ —46.5; H ₂ O—14.0.
Labradorite: 2.71—Ab ₂ An ₂ : Na ₂ O—4.6; CaO—12.3; Al ₂ O ₃ —30.0; SiO ₂ —53.1.
Leucite: 2.48—KAlSi ₃ O ₈ : K ₂ O—21.5; Al ₂ O ₃ —23.3; SiO ₂ —55.2.
Limonite: 3.8—2Fe ₂ O ₃ ·3H ₂ O: Fe—59.8; O—25.7; H ₂ O—14.5.
Magnesite: 3.04—MgCO ₃ : MgO—47.6; CO ₂ —52.4.
Magnetite: 5.13—Fe ₃ O ₄ : FeO—31.0; Fe ₂ O ₃ —69.0 (or Fe—72.4; O—27.6).
Melilite: 3.0—Na ₂ (Ca, Mg) ₁₁ (Al, Fe) ₄ (SiO ₄) ₃ : Na ₂ O—4.3; CaO—31.3; MgO—8.4; Fe ₂ O ₃ —11.2; Al ₂ O ₃ —7.1; SiO ₂ —37.7.
Muscovite: 2.87—(OH) ₂ KAl ₃ Si ₃ O ₁₀ : K ₂ O—11.8; Al ₂ O ₃ —38.5; SiO ₂ —45.2; H ₂ O—4.5.
Nephelite: 2.6—K ₂ Na ₄ Al ₃ Si ₃ O ₁₄ : K ₂ O—7.7; Na ₂ O—15.1; Al ₂ O ₃ —33.2; SiO ₂ —44.0.
Oligoclase: 2.66—Ab ₁ An ₁ : Na ₂ O—8.8; CaO—5.2; Al ₂ O ₃ —23.9; SiO ₂ —62.1.
Opal: 2.1—SiO ₂ ·xH ₂ O: SiO ₂ —85 to 97; H ₂ O—3 to 12.
Orthoclase: 2.56—KAlSi ₃ O ₈ : K ₂ O—16.9; Al ₂ O ₃ —18.4; SiO ₂ —64.7.
Pyrite: 5.03—FeS ₂ : Fe—46.6; S—53.4.
Pyrrhotite: 4.59—Fe ₁₁ S ₁₂ : Fe—61.5; S—38.5.
Quartz: 2.66—SiO ₂ : SiO ₂ —100.
Rutile (etc.): 4.2—TiO ₂ : TiO ₂ —100.
Serpentine: 2.56—H ₂ Mg ₃ Si ₂ O ₅ : MgO—43.0; SiO ₂ —44.1; H ₂ O—12.9.
Siderite: 3.86—FeCO ₃ : FeO—62.1; CO ₂ —37.9.

Sodalite: 2.3— $\text{Na}_4\text{Al}_3\text{Cl}(\text{SiO}_4)_3$: Na_2O —25.6; Cl —7.3; Al_2O_3 —31.6; SiO_2 —37.2.
Staurolite: 3.70— $\text{HFeAl}_3\text{Si}_3\text{O}_{11}$: FeO —15.8; Al_2O_3 —55.9; SiO_2 —26.3; H_2O —2.0.
Talc: 2.73— $(\text{OH})_2\text{Mg}_3\text{Si}_4\text{O}_{10}$: MgO —31.7; SiO_2 —63.5; H_2O —4.8.
Titanite: 3.52— CaTiSiO_5 : CaO —28.6; TiO_2 —40.8; SiO_2 —31.7.
Tourmaline: 3.1— $\text{R}_2\text{Al}_2\text{B}_2(\text{OH})_2\text{Si}_4\text{O}_{10}$: SiO_2 —35 \pm ; B_2O_3 —10 \pm ; Al_2O_3 —30 to 43; $(\text{FeO}.\text{MgO})$ —15 \pm ; Misc.—5 \pm .
Wollastonite: 2.85— CaSiO_3 : CaO —48.3; SiO_2 —51.7.
Zircon: 4.69— ZrSiO_4 : ZrO_2 —67.2; SiO_2 —32.8.
Zoisite: 3.31— $\text{HCa}_2\text{Al}_2\text{Si}_2\text{O}_{11}$: CaO —24.6; Al_2O_3 —33.7; SiO_2 —39.7; H_2O —2.0.

Several of the great groups of common rock-forming minerals are isomorphous and complex series, with great variety in their extreme phases. Chief among these are the feldspars, ranging from 68 to 43 per cent in silica. Because of their great abundance and wide range the differences in the feldspars are largely used in petrography, and many methods have been developed for their distinction (see pages 468 and 469). There seems to be a tendency among the most active workers to depend on the determination of indices of refraction in oils as the most rapid and reliable method when material is available. Similarly, other isomorphous groups are largely studied by immersion of the powder in oils.

The feldspathoids constitute a group of rather loosely related minerals, resembling feldspars in some respects but commonly having a lower percentage of silica. Nephelite and leucite are the most common, but the group includes melilite, analcite, cancrinite, scapolite, and the sodalite group. The word feldspathoid is long, and 'lenad' has been suggested as a short term for the same meaning, derived mnemonically from leucite (le) and nephelite (n); 'foids' is another abbreviation of similar significance; 'feloids' comprises both the feldspars and the feldspathoids.

The physicochemical mechanism of isomorphous replacement is still a matter of discussion. It seems that commonly the molecular volumes of the elements or groups in isomorphous mixtures are closely similar, but some series involve two end members with different molecular volumes, as for example the series from NaAlSiO_4 to KAlSiO_4 ; and in this series there is a further possibility of considerable additions of $\text{CaAl}_2(\text{SiO}_4)_2$, so that the relations are very complex.

The curves of variation in optical properties of important isomorphous rock-making minerals are given in Winchell's "Optical Mineralogy."

PETROGRAPHIC PROBLEMS

Problem 1. To sample a rock outcrop.

Method.—Collections differ not only with the nature of the material but also with the purpose of the collector. (1) For a display or official collection hand specimens can be trimmed 3 by 4 by 1 inches on an average, and certain special phases may require larger pieces. (2) For examination and classification a rough lump and chip for thin section may suffice. (3) For a research on the origin of some rock, for an analysis of variable bedded rocks, or for some special feature it may be necessary to take a series of specimens or a channel sample across the formation. Friable materials and earths that can not be trimmed to make per-

manent hand specimens may be sampled by cutting a channel or by the use of an auger, and the samples kept in bags, boxes or bottles.

Large samples, too heavy to be transported, need to be "quartered down." To quarter a sample accurately the pieces should be broken fine and heaped into a conical pile by pouring the material straight down on the top of the cone. The conical pile is then spread symmetrically into a flat round sheet of material which is marked off into four quarters by lines crossing in the center. Opposite quarters are saved to make a sample half the original size, and the alternate quarters are discarded. The process is repeated until the desired size is obtained. Care must be taken to crush to smaller fragments as the total sample becomes smaller.

The economic geologists and petrographers have in general developed different attitudes and different methods in sampling materials. A sample of an ore deposit is commonly taken in such a manner as to represent as nearly as convenient a fair average of a large tonnage of material in the ground. On the contrary, a petrographer, having found an interesting local rock phase or even an unknown volume of rock in the field, may find it desirable for description and classification to have specimens of that small mass. A few chips from a hand specimen serve for analysis and thin section. Much can be said in defense of each method, but the two methods are not equally valuable for all purposes.

The petrographer's method gives him data for a comparison of analysis and thin section of the same specimen. The chemical work can be checked by chips from the same specimen, if the results are questioned at a future time. If the analysis had been made on material collected as a series of chips from a large area, crushed and quartered to reasonable size, the resulting analysis might have been different from any chip in the set; it would have been an analysis of an average sample, not of any particular phase of the rock. As it is an actual rock that the petrographer is usually interested in, the average of a non-uniform mass, constituting a commercial tonnage, would have little value from his standpoint. To be sure, he is interested in a series of phases, and a complete study may involve testing a large number of specimens. Too often analyses have been made on a single, poorly selected specimen; but a single *well-selected* specimen of a fairly uniform mass has some advantages over a composite sample. A few grams of *fresh* rock are better than hundreds of pounds of somewhat *weathered* rock carefully quartered down.

The kind of problem to be solved may well modify the attitude to be taken in such sampling. A study of the serial variation in a single mass is probably best conducted by tests on specimens in the series as distinct from samples of great areas.

Problem 2. To make a thin section.

Method.—Work is started with chips about an inch square and as thin as consistent with strength. These are ground smooth by one of the several methods, for example, rotating lead laps with emery and water. Field outfits provide simply carborundum bricks. The smoothing moves from coarse to finer abrasives. Polishing is done on wet glass. The smooth surface is fastened to glass with heated Canada balsam, and the rock is ground thin (less than 0.05 mm.; the standard is 0.03 mm.). Commonly sections of hard rock are transferred to their final mount after

grading. Various precautions and details are best learned by experience.¹ Unconsolidated rocks need to be cemented by cooking in balsam or by impregnating with xylol balsam or bakelite varnish. Coal may require a preliminary treatment to make the organic material transparent or to soften it so that it can be cut² (see also the notes on mounting loose grains, Problem 4).

Problem 3. To make a mechanical analysis of loose grains.

Method. (Applicable chiefly to sediments that are not too well cemented.)—Shake 10 grams with water in an 8-ounce bottle for several hours. The disintegration of some specimens is assisted by boiling or by adding a few drops of ammonia, Na_2CO_3 or NaOH .

Especially tenacious aggregates may require hot or concentrated reagents or gentle rubbing with a rubber pestle. The use of acids destroys the calcareous fossils, if any are present. Bituminous matter in sediments may be extracted by benzol or chloroform before other treatment. Pyrite and barite, if abundant, may make a special treatment necessary; nitric or concentrated sulphuric acid may be used, but either is likely to attack several other minerals.

Sand more than $\frac{1}{32}$ mm. in diameter settles 15 inches in a cylinder of water in 5 minutes, and the silt and clay suspension can then be siphoned off into a beaker, Fig. 1. The silt $\frac{1}{32}$ to $\frac{1}{64}$ mm. in diameter will then settle 5 inches in 5 minutes, and the clay can be siphoned off. The several grades of sand may be separated by screens after the sand is dry. The clay portion is not usually separated further but may be centrifuged. All separates are then ready for further examination, which may be mainly mineralogic, textural or paleontologic. Usually in sediments all three kinds of examination yield results worthy of record.

Various workers have modified the treatment, some preferring a process of panning, others elutriation in a series of more complex forms of apparatus; the simpler methods are usually to be preferred.³

The sizes obtained by any sieve or customary treatment should be checked by micrometer measurement. Some workers in sediments have used other standards of sizes.

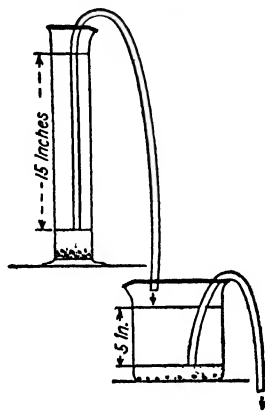


FIG. 1.—An arrangement of apparatus for the separation of sand grains of different sizes.

¹ KEYES, MARY G., Making thin sections of rocks, *Am. Jour. Sci.*, 5th ser., vol. 10, p. 538, 1925. C. S. ROSS, Methods of preparation of sedimentary materials for study, *Econ. Geology*, vol. 21, p. 461, 1926.

² WILLIAMS, T. B., Identification of coals, *Econ. Geology*, vol. 21, p. 364, 1926. REINHARDT THIESSEN, Structure in Paleozoic bituminous coals, *Bur. Mines Bull.* 117, pp. 10-13, 1920.

³ TICKELL, F. G., "The Examination of Fragmental Rocks," Stanford Univ. Press, 1931.

The examination of the separates depends on their size. Many samples of coarse sand may be identified without a microscope. Finer grains are often separated by the methods of Problem 4 and examined by immersion in oils. The clay is seldom determined mineralogically; chemical tests may be made on it.

Problem 4. To separate the minerals of a rock powder.

Methods.—All start with powder crushed to a size of grain smaller than the smallest grains it is hoped to separate. One of the best materials on which to use the methods is the sand or silt obtained as in Problem 3. Crushing of hard rocks is done by tapping rather than rubbing or grinding under a pestle. If much fine powder results from crushing (or is present in the original sample) it is best washed off, as the proportion in the residue is not much changed. A correction may be applied, because brittle minerals, like quartz, form more dust than ferromagnesian minerals.

After applying any of the following methods permanent mounts of the separates, especially of the heavy residuals of sediments, are commonly made (1) in air, so that the frosted or polished surface of grains can be studied; a perforated card on the slide serves to hold the grains under a cover glass, and for some examinations a black card is preferred to a glass slide; (2) in balsam with an index about 1.54; (3) in piperine, index about 1.65; and (4) in hyrax, index about 1.8; the three media of different index help greatly in identifying the minerals.

Method 1.—The powder from any separate obtained as in Problem 3 or the later methods of Problem 4 should be inspected before selection of further treatment. If the material is a fairly coarse sand and only one or two grains of some heavy or characteristic mineral are present, it may be best to pick out the grains with a needle or with a small rod dipped in balsam. Xylol will later remove the balsam from the grains. Grains too large to give good results in oils may be crushed between two stout slides; but even a single grain, if of convenient size, may be optically studied by changing the immersion oils and by rotating the grain in each oil.

Method 2.—The method most recommended in igneous petrography is the use of a heavy liquid. (a) Thoulet solutions KHgI_3 may be concentrated to a gravity of 3.1. Use 5 parts of HgI_2 , 4 parts of KI and 1 part water. (b) AgNO_3 . TiNO_3 is a liquid at 75°C . with a gravity of 4.5. (c) Clerici solution; thallium malonate and formate can be concentrated to a specific gravity of 4.95. These can be reduced by dilution with water. (d) Methylene iodide has a gravity of 3.3 and can be reduced with benzol. Very few minerals except sulphides are attacked by these solutions, so that by adjusting the gravity to a figure between those of two rock minerals a separation can easily be effected. By successive dilutions a series of minerals can be settled out. Small lumps of known specific

gravity may be useful as indicators. (e) For the separation of sediments the use of bromoform is recommended. It can be used with a gravity of 2.90, and a further separation can be made with a dilution to 2.69 by means of acetone. *Keep acetone away from the flame.* The same liquid is used for washing. Commercial bromoform commonly needs purification by washing with excess of water and separation in a separatory funnel. If diluted with acetone it can be recovered by shaking with a large excess of water, which dissolves the acetone, and again separating in a funnel.

Several types of apparatus are arranged for convenience in separating powders. A simple funnel with a stopcock at the outlet usually proves satisfactory, especially if it is desired to separate a few heavy grains from a large mass of light material. If quantitative work is not needed, a preliminary concentration may be accomplished in a beaker by skimming off the grains that float. For careful separation of grains that make up large portions of the rock a Penfield separator is very useful (Fig. 2).

Method 3.—Heavy minerals can be separated from light ones by repeated panning, as in panning for gold. The method is most used for separating small quantities of heavy accessory minerals from a large amount of rock and works best if there is a large difference in gravity. Some workers prefer a conical pan to the common miner's pan.

Method 4.—The several minerals of a rock are usually attracted to different degrees by a magnet. The magnetic permeability of common minerals is somewhat variable, but it may be possible to separate into as many as six groups. The fractions obtained should be cleaned by repeating the separation.

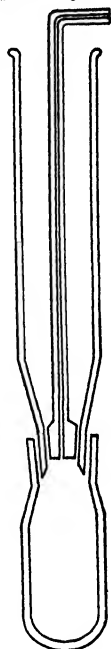


FIG. 2.—
Penfield apparatus for separating minerals in heavy liquids.

- | | |
|---------------------|---------------|
| 1. Metallic iron | Glauconite |
| Pyrrhotite | Pyrope |
| Magnetite | Xenotime |
| | Monazite |
| 2. Ilmenite | 4. Tourmaline |
| Wolframite | Bronzite |
| Franklinite | Vesuvianite |
| Chromite | Cassiterite |
| Hematite | Columbite |
| Limonite | |
| Siderite | 5. Staurolite |
| Augite (rich in Fe) | Actinolite |
| Spinel (rich in Fe) | Olivine |
| Almandite | Pyrite |
| Arfvedsonite | Pyrope |
| Lepidomelane | |
| Manganese oxides | 6. Biotite |
| Alabandite | Chlorite |
| | Rutile |
| 3. Hornblende | Quartz |
| Augite (poor in Fe) | Feldspar |
| Epidote | |

Method 5.—Electrostatic separation may be effected as follows: Spread the dry powdered rock on a sheet of metal and bring over the powder a flattened stick of sealing wax excited by rubbing on cloth. Good conductors are attracted.

Method 6.—Soluble minerals may be dissolved away from insoluble ones.

To remove the calcite from an altered rock, use acetic acid.

To remove olivine and other soluble minerals from common feldspars and insoluble minerals, digest the powder in 1:1 HCl in large volume.

To separate kaolin from quartz and orthoclase in a potter's clay, decompose the kaolin by 1:1 sulphuric acid, boiled until it fumes. Dilute, decant the liquid, and finish the solution by warming a few minutes with 5 per cent Na_2CO_3 .

To separate the insoluble residues from a limestone or dolomite, use a large volume of dilute hydrochloric acid. See page 264. Use caution in heating it, for cooling may result in some deposits of gypsum.

To separate nephelinite from olivine; it is said that boiling with 1:40 nitric acid for one minute will dissolve nephelinite but not olivine. It is well, however, to test the solution for the presence of iron, which would indicate that the olivine was somewhat attacked.

Several minerals can be separated from a silicate rock by their dissolving in HF more slowly than most minerals. Even combinations of HF and H_2SO_4 attack them very slowly. The most common minerals of the group are magnetite, hematite, zircon, rutile, corundum, tourmaline and staurolite. The rarer insoluble minerals include anatase, brookite, perovskite, andalusite, sillimanite, kyanite, cassiterite, topaz, axinite, and spinel.

Method 7.—The commercial process of oil flotation may be applied to rock minerals, but good laboratory methods for rock separation by this process have not been developed.

Many small fossils containing air cells may be floated from fine sediments by throwing the powder dried at 150° on hot water.

Method 8.—Mica may be separated by its scales in a powder sticking to an inclined paper moistened beforehand by breathing on it.

NOTE.—Combinations of methods are especially attractive. After a mechanical analysis of a sediment, or a separation by heavy liquid, the fractions may be divided by magnetic and electrostatic methods with excellent results. After magnetite is removed from an olivine gabbro, the olivine may be dissolved from the feldspar. Other variations are too numerous to mention.

Problem 5. To test a rock for minerals that give a silica jelly.

Method.—Boil a gram or more of fine powder thoroughly with 1:1 HCl. Filter or decant and boil down the solution. If the soluble silicate was abundant, the jelly may form during evaporation. If not, boil down to two or three drops and set aside to cool quietly for ten minutes, when even small amounts of jelly can be detected. Distinguish a jelly from a thick syrup.

The chief rock minerals that give a silica jelly are nephelinite, sodalite, olivine, chondrodite, anorthite, noselite and cancrinite.

Less definitely and less uniformly a jelly may be obtained from analcite, leucite, melilite, meionite, wollastonite, serpentine, chlorite, glasses and other zeolites.

Problem 6. To make a chemical analysis of a rock.

Method.—Refer to H. S. Washington, "Rock Analysis" (4th ed., John Wiley & Sons, Inc., 1930); W. F. Hillebrand, The analysis of silicate and carbonate rocks (U. S. Geol. Survey Bull. 700); and Hillebrand and Lundell, "Applied Inorganic Analysis" (John Wiley & Sons, Inc., 1929). Estimate from microscopic examination what accessory elements a chemist should search for in several igneous rocks. Minute traces of zircon and titanite are more easily estimated from section than by analysis.

Problem 7. To estimate the analysis of a rock by a microscopic examination.

Method.—The first step is an estimation of the proportion of minerals present—the mode. Using a magnification high enough to recognize all the minerals and a micrometer eyepiece (or mechanical stage), a line is selected at random across the section, and for each mineral that the line crosses record is made of the length of the crossing. If the rock is coarse, a series of parallel lines at regular intervals are so recorded. The sums of the lengths recorded for each mineral may be assumed to be proportional to the volumetrical occurrence of the minerals in the rock. This is the first of several assumptions involved. The accuracy of the method increases with the uniformity of the rock, the size of the thin section, and the length of the line measured. A correction for the specific gravity, and recalculation to 100 per cent give the per cent of each—a mineral analysis. A convenient form for calculation is shown by Pirsson.¹ The results are commonly called a Rosiwal analysis.

This is translated into chemical terms by using an assumed analysis of each mineral measured (see page 10). The selection is commonly an average of analyses such as are recorded in Dana, "System of Mineralogy"; Iddings, "Igneous Rocks," vol. 1, p. 454; Osann, "Beiträge zur chemischen Petrographie," Part II, supplement 1905; Doelter, "Handbuch der Mineralchemie." The feldspar is determined as on page 468, but the method is liable to serious errors in the estimates of soda and potash. The minerals of most variable composition aside from feldspars are the amphiboles, pyroxenes and micas. In augite, for example, silica ranges from 44 to 52 per cent, and ferrous oxide from 2 to 12 per cent. Some varieties have a characteristic color that indicates composition, but the result is always uncertain if these materials make up a large part of the rock. (For a discussion of methods, see Lincoln and Rietz, Econ. Geology, vol. 8, pp. 120–139, 1913.) If possible to test some of the pow-

¹ PIRSSON, L. V., Igneous rocks of the Highwood Mountains, U. S. Geol. Survey Bull. 237, p. 73, 1905.

dered mineral in oils for the indices of refraction, the estimated composition may be made more accurate. See Winchell, "Optical Mineralogy."

After sketching with a camera lucida (Problem 9), an estimated mineral analysis may be made by the use of a planimeter, or by cutting out and weighing the paper represented by each mineral. A microprojection may be even more serviceable than the camera lucida.

Problem 8. Petrographic calculations.

These are based largely on analyses (or calculated analyses, Problem 7) but may involve a variety of other data many of them more or less inaccurate (pages 148 to 158). The student should realize fully that no amount of calculation can give reliable results if based on inaccurate data or unjustified assumptions. The calculations may be needed, even if the data are not entirely accurate, and may serve to place limits within which the true values lie. For this work the mathematicians have methods of keeping track of the probable errors,¹ but the original data should be judged with care before important conclusions are based upon them.

A. Given the specific gravities of the minerals of a rock and their proportions by weight or by volume, what is the specific gravity of the rock?

Method.—The specific gravity of a rock, assuming its substance to be non-porous, may be calculated by the formula:

$$D = \frac{v'd' + v''d'' + v'''d''' \dots}{100}$$

or

$$D = \frac{100}{\frac{w'}{d'} + \frac{w''}{d''} + \frac{w'''}{d'''} \dots}$$

where D is the required aggregate specific gravity; v' , v'' , v''' , etc., are the volume percentages of constituents I, II, III etc.; w' , w'' , w''' , etc., are the weight percentages; and d' , d'' , d''' , etc., are the specific gravities of the same series of constituents.²

Some have also used the specific gravities calculated from normative minerals, instead of from the mode.³

B. Relations of porosity, volume, and specific gravity. Most rocks are porous to some extent, and the specific gravity may be determined

¹ General references for this problem are:

HOLMES, A., "Petrographic Methods and Calculations," Murby & Co., 1920; and F. F. GROUT, The use of calculations in petrology, Jour. Geology, vol. 34, pp. 512-558, 1926. Rules for keeping track of errors are given in the latter paper, pp. 522-524.

² SHARWOOD, W. J., The specific gravity of mixtures, Econ. Geology, vol. 7, pp. 588-590; 1912.

³ IDDINGS, J. P., Densities of igneous rocks calculated from their norms, Am. Jour. Sci., 4th ser., vol. 49, pp. 363-366, 1920. See also Washington's estimates of specific gravity in Franklin Inst. Jour., vol. 190, pp. 804-806, 1920.

with or without the included pores. Absorption is commonly estimated by boiling four hours in water, cooling in water, and noting the increase in weight.

$$\text{Porosity} = \frac{\text{water absorption in grams}}{\text{volume in cubic centimeters}}$$

The true specific gravity taken in the absence of pores, multiplied by (100 — per cent porosity) gives one-hundred times the “apparent” specific gravity, taken on a specimen with pores included; or, transposing,

$$\text{Per cent porosity} = \frac{\text{true specific gravity} - \text{apparent specific gravity}}{\text{true specific gravity}} (100)$$

C. Volume changes in metamorphism are calculated by the following rule: The volume of the original compound is to the volume of the compound produced directly as their molecular weights and indirectly as their specific gravities.

D. Atomic and molecular proportions. One further simple calculation from data leads to the more complex schemes of calculation. This is the problem of atomic proportions or molecular proportions. A general familiarity with the idea involved serves many purposes, but a number of cases have been found in which a student, even after making the elaborate calculations of a norm, had no understanding of the meaning of molecular proportions.

Suppose a chemist reports the composition of a sulphide as 47 ± 1 per cent Fe and 53 ± 1 per cent S. What is the probable formula of the mineral? The atomic weights are: Fe = 55.84 and S = 32.064. The errors in chemical data are about 2 per cent of the amounts reported, so that the data need not be elaborated farther than to keep three significant figures.

The algebraic rule is to let x be the number of atoms of sulphur for each atom of iron. Then

$$\frac{55.8}{32.0x} = \frac{47 \pm 1}{53 \pm 1} \text{ and } x = 1.96 \pm 4.$$

More generally use is made of the arithmetical rule: Divide the percentages by the atomic weights to get the atomic ratios.

$$\frac{47 \pm 1}{55.8} \text{ atoms of Fe combine with } \frac{53 \pm 1}{32.0} \text{ atoms of S}$$

or

$$0.84 \text{ atom of Fe combines with } 1.65 \pm 3 \text{ atoms of S}$$

The formula should be FeS_2 , pending more accurate chemical work.

The next stage in the complexity of such calculations involves the common custom of reporting rock analyses in terms of oxides rather than elements. Instead of determining atomic ratios in a feldspar, for example,

one estimates molecular ratios of K_2O , Al_2O_3 , SiO_2 , etc., as if feldspars were made up of oxides. The fallacy of the idea should not mislead anyone, and the convenience of the method makes it widely used.

E. To calculate the standard mineral composition—the *norm*—from a chemical analysis and classify a rock in the quantitative chemical system. Follow the directions on page 188 *et seq.* of the Quantitative mineralogical-chemical classification by Cross, Iddings, Pirsson, and Washington, or page 1163 *et seq.* of Professional Paper 99, the U. S. Geological Survey. Use the tables instead of calculating. Try first to duplicate some example given in the book. Second, use the analysis that you have estimated in Problem 7 and see the difference between the norm calculated and the mode measured. See also the CIPW book, pages 226 to 230.

F. To calculate the *mode*, or actual mineral analysis, from the list of minerals known to be present and the chemical analysis.

Method.—Calculate the fixed molecules as before. If there are among the less definite molecules more variables than equations it is necessary to make some assumptions—more or less accurate.¹ Type rocks have been calculated for comparison.²

Try out Mead's geologists' slide rule and Hance's tables (page 10) with a slide rule instead of making calculations in detail or using the tables.³

Aurousseau⁴ suggests a detailed tabulation of a mode and analysis showing not only what oxides are present and how much, but *where they are* in the mode minerals.

G. To classify a rock in other schemes. See Johannsen, "A Descriptive Petrography of the Igneous Rocks," vol. 1, pages 59–158, University of Chicago Press, 1931.

Several of these calculations refer to "standard" minerals as contrasted with actual minerals of rocks. These arbitrarily adopted minerals are so much criticized that most petrographers hesitate to adopt them as a basis of classification. Rocks are said to be chemically characterized in the common classifications as well as in those specially based on analyses.⁵

H. To calculate the additions and subtractions during rock alteration. The difficulty of knowing exactly what happened in an alteration is shown by Lindgren in two examples,⁶ namely, the alteration of pyrargyrite to argentite and the alteration of olivine to serpentine.

¹ KEMP, J. F., "Handbook of Rocks," Chap. 13, D. Van Nostrand Company, Inc.

² For examples and types see U. S. Geol. Survey Prof. Paper 76, pp. 105, 106, 108, 110, 129–130, 141–143, and 152–154.

³ MEAD, W. J., Some geologic short-cuts, Econ. Geology, vol. 7, pp. 136–144, 1912; J. H. HANCE, Use of the slide rule in the computation of rock analyses, Jour. Geology, vol. 23, pp. 560–568.

⁴ New South Wales Linnean Soc. Proc., vol. 51, part 4, pp. 622–623, 1926.

⁵ MATHEWS, E. B., Chemical characterization of rock types, Geol. Soc. America Bull., vol. 41, pp. 56–57, 1930.

⁶ LINDGREN, WALDEMAR, Metasomatic processes in fissure veins, Am. Inst. Min. Eng. Trans., vol. 30, pp. 578–692, 1900.

The problems become definite only when we know the volume of the secondary substance, or when we know that some constituent remains constant.

Case 1. *The Method of Constant Constituent.*—The method of assuming a constant constituent is used largely with weathered rocks in many of which it can be shown that alumina is not dissolved appreciably. Lindgren gives the method and rules and formulas in sufficient detail.¹ As a whole, however, emphasis should be put on Lindgren's warning that the assumption of a constant constituent should be made only with the greatest caution.

Case 2. *The Method of Constant Volume.*—The volume of many altered rocks is no doubt approximately that of the equivalent fresh rock. This has been assumed for many specimens from the slender evidence of porosity and texture; it has been strongly maintained by Lindgren; and the method is justified by recent accurate studies at Bingham.² For accurate work there should be available not only the analyses of fresh and altered rocks but also their specific gravities in powder and in bulk (or a porosity determination with one of these).

Calculate first the quantities of each constituent in 100 cc. of fresh rock and then the same in altered rock. Comparison shows the gains or losses, which may then be transformed into percentage of the original rock, or percentage of the original amount of that constituent present.

Case 3. *The Method When No Constant Can Be Assumed.*—The analyses of fresh and altered rocks may still be studied with some profit by graphic methods on a "straight line" diagram. See Fig. 256.

J. Differentiation and assimilation. The calculations are often made by the method of successive approximations (see the general references at the beginning of Problem 8).

Problem 9. To diagram the nature of rocks.

Method.—The quantitative data as to the minerals present, their forms, their densities, their optical properties, and data as to the chemical constituents may be plotted either alone or on coordinate systems. Perhaps the most common and valuable plot shows these features as abscissae, while the ordinates represent the positions of the samples in the field (see Fig. 109; also Holmes, "Petrographic Methods and Calculations," pages 440 to 491). The series may be a result of differentiation, alteration, or sedimentation.

The variations in several series of igneous rocks may be shown on a single plot, using silica and alkali contents as coordinates (Fig. 111).

The characters of sedimentary rocks, abundance of the several minerals, sizes of grain, etc., are commonly shown by bars of different lengths in a composite diagram for each rock (see page 264 and Fig. 165).

¹ "Mineral Deposits," Chap. 20, 1913.

² LINDGREN, WALDEMAR, Contact metamorphism at Bingham, Utah, Geol. Soc. America Bull., vol. 35, pp. 507-534, 1924.

Problem 10. To illustrate a rock description.

Method 1.—Make free-hand sketches.

Method 2.—Sketch with a camera lucida. Attach the apparatus to the microscope at the ocular. Arrange the drawing paper at a level near that of the stage of the microscope. Insert the diaphragm that equalizes the illumination from the microscope and the paper. Part of the sketch may be made with nicol prisms and part without.

Method 3.—Use a camera with a ground glass for focusing and place in alignment before the microscope ocular. Avoid letting any side light enter the camera. Time the exposure according to the light used. The timing may need experiment.

Use also the special apparatus for photomicrography.

If a section contains two opaque ores of different color, they may be shown by combining reflected and transmitted light.

India ink lines on a photograph may be preserved as a line drawing after a treatment that dissolves away the picture.¹

Problem 11. To write a rock description.

Methods depend on the purpose of the work, whether (1) for classification and naming, (2) for commercial information, or (3) for a complete record of a formation. It should be added that peculiar structures and textures are better depicted by a sketch or photograph than by whole pages of text (see Problem 10 and Figs. 34, 40, 63, 79, 80, 190, 194, 217, 218, 225, etc.).

Method 1.—For classification, naming and applied problems petrographic blanks are very useful and emphasis is easily put on the particular feature needed (page 24). The history of events in the development of a rock is of the utmost value in many commercial problems.

Method 2.—For a complete descriptive record a similar blank may be used, but whether it is or not care should be used to cover the following points.

1. Location, commonly paced from some point located by a government survey.
2. Field relations and structures.
3. Description of the hand specimen, its color and tone, and the impression it makes.
4. Texture.
5. A list of the minerals, preferably accompanied by a brief statement as to which are essential, or a record of percentages of the important constituents, as nearly quantitative as possible.
6. A description of the important minerals, avoiding useless detail. Only those characters of minerals should be mentioned that serve to show how the rock being described differs from other rocks; such features as size, form, twinning peculiarities, pleochroism, color, inclusions, and anomalous characters may be of some value. The sequence of crystallization of the minerals is inferred largely from the forms and textures.

¹ NAYLOR, E., Pen and ink drawings from photographs, *Science*, vol. 73, p. 15, 1931.

7. An analysis, if available; and calculations based upon it.
8. A descriptive name, with some remarks on the reasons for such classification and the origin and the history inferred.

Sample descriptions in great detail for educational use are published in Bulletin 150 of the U. S. Geological Survey.

One of the most detailed studies chemically is by Aurousseau, in the paper cited (problem 8). He tabulates not only the amounts of the oxides present, but also the minerals in which they occur.

Method 3.—In many reports a petrographer should describe a *formation* rather than a few rock specimens, and the outline above should be generalized from particular specimens and incorporated as the “petrographic character” in a large outline such as the one below. Good examples have been given by Clapp¹ and by Bliss and Jonas.²

Distribution.
Field and structural relations.
Petrographic character.
Normal rock.
Variations.
Lateral and vertical.
Border and transitional.
Alteration.
Contact action.
Thickness.
Origin.
Name of the formation.

SAMPLE ROCK DESCRIPTIONS

No. 1. For an Engineering Report³

Granite.—The rock is coarse grained and well interlocked. The chief constituents are orthoclase, quartz and microcline.

There are but small amounts of dark minerals, and there is not much decay.

Both surface material and the drill core were examined. The deeper material shows a little calcite, that may be original, occurring in irregular grains. They do not seem to indicate decay. There is a little kaolin alteration of the feldspars, but not to a serious degree. There are no injurious impurities in the rock such as might cause rapid disintegration or discoloration.

The rock is undoubtedly of good grade as to strength, composition, and durability

No. 2. A Brief Petrographic Report⁴

Syenite.—The early porphyritic augite-syenite is characterized by the abundant and consistently rounded pink and light-gray feldspar phenocrysts, which average about one-half inch in length. The ground mass is fine and contains augite, hornblende and less abundantly biotite. In a typical specimen, taken from the locality where the sill-like relations of the rock are suggested the mineral composition is as follows: soda-potash feldspars, 80 per cent; augite, 13 per cent; hornblende, 3 per cent; biotite, 2 per cent; accessory titanite, magnetite and apatite. The augite has a

¹ CLAPP, C. H., Geology of the igneous rocks of Essex County, Mass., U. S. Geol. Survey Bull. 704, 1921.

² BLISS, E. F., and A. I. JONAS, U. S. Geol. Survey Prof. Paper 98, pp. 9-34, 1917.

³ BERKEY, C. P., Geology of the New York City aqueduct, New York State Mus. Bull. 146, p. 198.

⁴ SANDERS, O. W., Geology and petrography of the Snowbank composite Stock of Minnesota, Jour. Geology, vol. 37, No. 2.

tendency toward aegirite-augite. Soda in the molecule, probably up to 2 or 3 per cent, is suggested by the following considerations: (1) by the pleochroism, which is $X =$ light grayish green, $Y =$ light grayish green, $Z =$ yellowish green, absorption: $X = Y > Z$; (2) by the extinction angle, which attains a maximum of only 38 degrees; (3) by analogy: the augite from the granite of Kekequabic Lake (in the same petrographic province) shows approximately the same pleochroism, and its maximum extinction is 37 degrees. It has 2.63 per cent Na_2O .

Hornblende and biotite have been formed apparently by the action of the late magma on the pyroxene. In some of the specimens epidote is abundant among the secondary minerals.

No. 3. For a Mining Report

PETROGRAPHIC DESCRIPTION (Blanks by C. P. Berkey)

Date 1929

Collector F. Grout

I. FIELD NOTES

Original No. GP 926 Locality: Lunsemfwa, Rhodesia
Occurrence: Prospect pits in granite lenses in gray gneiss
Question: Is the chalcopyrite primary or introduced?

II. HAND SPECIMEN DESCRIPTION

General Appearance: Pink, aplitic, banded faintly; with chalcopyrite and quartz

III. MICROSCOPIC STUDY FOR CLASSIFICATION

Texture: Granitoid
Original Structure: Massive (to trachitoid?)
Secondary Structure: Very slightly altered

MINERALOGY (Minerals are grouped for interpretation purposes and are arranged in each group in approximate order of abundance)
(In some cases approximate percentages are given)

PRIMARY (X) Essential Minerals	%	(Z) SECONDARY Alteration Products	%	(M) METAMORPHIC Recrystallization Minerals	%	(T) TERTIARY CHANGES AND ENRICHMENT EFFECTS
Microcline Orthoclase Quartz Plagioclase		Chlorite Sericite Leucoxene-rutile				Tarnish (Bornite?)
(Y) Accessory Minerals		(G) NATURE OF GROUNDMASS OR CEMENT		(O) INTRODUCED SUBSTANCES OR MINERALIZATION		
Apatite Biotite Zircon Rutile Tourmaline?				Chalcopyrite		

SPECIAL FEATURES:

Irregular apatite, corroded?

Chalcopyrite occurs definitely between and crossing the quartzes and in veinlets, replacing it a little.

The sericitization is so slight that it does not seem likely that the sulphide was hydrothermal at any moderate depth. Very little alteration.

Might be considered an introduction and recrystallization at high temperatures, from a later intrusive, but this rock itself seems to be the latest intrusive. So the sulphides are probably deuteric.

CLASSIFICATION Chalcopyrite granite

ORIGIN OF THE ROCK: Probably deuteric introduction and replacement in a granite

PETROGRAPHER

F. Grout.

Fig. 3.—Reduced copy of a petrographic report for a mining company.

No. 4. *A Brief Petrographic Report*¹

Conglomerate.—Wy 50. Taken $\frac{1}{4}$ mile from mouth of tributary flowing westward from Hunt Mountain to South Beaver Creek, in float not far above granite; similar material in place 30 feet south and 10 feet above granite.

Description of hand specimen: Arkose conglomerate, pebbles not ranging above $\frac{3}{8}$ inch. Prevailing color where fresh dark-brown shot with gray, the pink of feldspar sharply contrasting; from dirty grays and whites of weathered surface pebbles stand out in relief. Fucoid markings on what may be bedding planes 1 inch apart. Pebbles 90 per cent quartz, faintly brownish-green, rounded to near-angular, breaking with matrix; feldspars pink fresh-cleavage fragments. Matrix 90 per cent of rock; sand, fine-grained, dull to earthy luster.

Texture granular-fragmental, large grains showing micrographic intergrowth. Constituents, quartz 90 per cent; microcline, orthoclase, uncertain plagioclase 8 per cent; small amounts of biotite, apatite, zircon as accessories, and sericite, kaolin, limonite and chlorite as alteration products. Liquid and gas inclusions in quartz. Cement a sericitic-kaolinic-limonitic "mess."

Quartz vari-sized, 1.5 by 0.8 mm. in larger grains, perhaps vein quartz, to judge by wavy extinction; average grains 0.18 mm.; rounded to near-angular. Microcline fresh, 0.03 to 0.04 mm. Orthoclase same size, largely sericitized. Much organic material, seemingly chitinous.

Classed: ferruginous arkose-conglomerate.

No. 5. *For an Economic Publication*²

Diorite Porphyry, Near Aspen.—The diorite porphyry is a dark fine-grained rock, and, as it is not obviously porphyritic, it resembles a finely granular diorite. Its most distinctive feature is that it is spotted black by sporadic phenocrysts of hornblende or its alteration products. It is a rock of marked individuality and is easily distinguishable from any of the other intrusive rocks of the district. On Richmond Hill it is liberally spotted with small areas of yellowish-green epidote, as well as by the dark spots of altered hornblende.

Under the microscope the diorite porphyry from Richmond Hill is seen to consist of feldspar phenocrysts, now completely altered to epidote except where rarely the peripheral zone of albite has remained intact, and ferromagnesian phenocrysts, largely chloritized, embedded in a microgranular groundmass of quartz and albite. The feldspar of the groundmass is fresh and clear, thus contrasting notably with that of the phenocrysts; it has evidently escaped alteration to epidote, like the peripheral zones of some of the plagioclase phenocrysts, because of its albitic, noncalcic composition. From the abundance of chlorite in the rock, the diorite porphyry was evidently a fairly basic variety, much more basic than any of the other igneous rocks in the district.

* No. 6. *For an Economic Publication*³

Granodiorite.—The dike $\frac{1}{8}$ mile southeast of the Cactus tunnel (specimen 110) is a gray rock containing rather abundant phenocrysts of feldspar with a much less number of biotite and one of the other dark silicates. Under the microscope the phenocrysts are seen to be oligoclase-andesine, biotite, and pseudomorphs of serpentine and chlorite after hornblende, and possibly pyroxene, in a fine groundmass of plagioclase, orthoclase and biotite, with chlorite and serpentine, which probably represents the alteration of hornblende and augite; a little quartz is also present. Iron ore is plentiful, and small crystals of apatite are abundant. The character of the plagioclase of the groundmass could not be definitely determined.

The proportion of the feldspars in the groundmass is also uncertain, but the phenocrysts are plagioclase. No fresh augite or hornblende is present, but there are perfect pseudomorphs after hornblende and irregular ones that may represent augite.

¹ TIEJE, A. J., Description and naming of sedimentary rocks, Jour. Geology, vol. 29, pp. 658-659, 1921.

² KNOPF, ADOLPH, Recent developments in the Aspen district, Colo., U. S. Geol. Survey Bull. 785, pp. 8-9, 1926.

³ BUTLER, B. S., Geology and ore deposits of the San Francisco and adjacent districts, Utah, U. S. Geol. Survey Prof. Paper 80, pp. 60-61, 1913.

Biotite is the most abundant of the ferromagnesian silicates. This is strongly pleochroic, being similar to that of quartz monzonite.

A partial analysis of specimen 110 by R. C. Wells gives the following composition:

PARTIAL ANALYSIS OF PORPHYRITIC DIKE ROCK FROM SAN FRANCISCO DISTRICT, UTAH

SiO ₂	61.52
CaO.....	2.91
MgO.....	1.19
K ₂ O.....	2.31
Na ₂ O.....	4.48

From the analysis the relative proportion of the principal minerals is calculated approximately as follows: quartz, 19; orthoclase, 12; oligoclase, 51 (albite, 38; anorthite, 13); biotite and hornblende, 7. The rock has the composition of granodiorite.

No. 7. *A Detailed Study for an Economic Report*¹

Quartz Monzonite.—Coarsely granular quartz monzonite occurs in large volume along the east flank of the Sierra Nevada. The type specimen was obtained on Lone Pine Creek at an altitude of 6,700 feet; but rock closely similar, if in fact not identical, is found at Lake Sebrina, South Lake, Division Creek, Shephard Creek, and Muah Mountain. In other words, it occurs at numerous localities and in extensive masses along the whole length of the part of the range examined.

The quartz monzonite is essentially an even-grained aggregate of plagioclase, orthoclase, quartz, biotite, and hornblende, and contains titanite in sporadic crystals. The grain is fairly coarse, ranging in the type specimen from 5 to 10 mm. The biotite and hornblende occur in roughly equal amounts and make up 15 per cent of the rock. They are in general closely associated and partly intergrown; the hornblende is nearly anhedral but the biotite tends to form hexagonal plates. The orthoclase is somewhat pinkish and is easily discriminable from the white striated plagioclase. The two feldspars are present in obviously equal quantities, and the quartz-monzonitic character of the rock is therefore readily apparent megascopically. Under the microscope a few additional features appear. The fabric is hypidiomorphic granular, owing principally to the subhedral development of the plagioclase. The order in which the different minerals commenced to crystallize is not certainly determinable, but quartz was the last to finish crystallizing. It is notable that small crystals of hornblende inclosed in the plagioclase are sharply euhedral, but that otherwise the hornblende is anhedral. The plagioclase has the composition Ab₆₅An₃₅. The orthoclase has been in part converted to microcline, and a small quantity of the curious intergrowth termed "myrmecite" is found; both of these newly developed structures evidently point to slight pressure effects. Titanite, magnetite, apatite, and zircon comprise the accessory minerals.

The chemical analysis and mineral composition of quartz monzonite from Lone Pine Creek is given below. In computing the mineral composition the lime, after deduction of the amount necessary for titanite and apatite, was distributed between the hornblende and plagioclase so as to produce a plagioclase of the composition Ab₈₀An₂₀, as indicated by the microscopic diagnosis.

ANALYSES OF QUARTZ MONZONITE FROM LONE PINE CREEK, CALIFORNIA
Chemical Composition

(R. C. Wells, analyst)

SiO ₂	66.68	H ₂ O+.....	0.51
Al ₂ O ₃	15.12	TiO ₂	0.64
Fe ₂ O ₃	1.66	CO ₂	None
FeO.....	2.49	P ₂ O ₅	0.21
MgO.....	1.38	S.....	Trace
CaO.....	4.09	MnO.....	0.01
Na ₂ O.....	2.23	SrO.....	Trace
K ₂ O.....	4.97		
H ₂ O—.....	0.31		
Specific gravity 2.715			100.30

¹ KNOPF, ADOLPH, A geologic reconnaissance of the Inyo Range, etc., U. S. Geol. Survey Prof. Paper 110. pp. 63-64, 1918.

Mineral Composition

Quartz.....	25.68
Orthoclase.....	25.58
Plagioclase (Ab ₆₀ An ₄₀).....	31.13
Biotite.....	7.02
Hornblende.....	7.38
Magnetite.....	1.39
Titanite.....	1.18
Apatite.....	0.31
	99.67

No. 8. A Detailed Complete Petrographic Record¹

Biotite-soda Granite Porphyry (Kallerudose, I.4.1.4).—Occurrence.—Biotite-soda granite porphyry forms the Marble Hill (Arizona) laccolith and is found only at the central peak.

Megascopic Characters.—The rock is light bluish gray in color and has a uniform aphanitic texture. Phenocrysts of prismatic feldspar and anhedral quartz (average length, 3 mm.; maximum, 8 mm.) form about 5 per cent of the rock; biotite 1 mm. in maximum length constitutes 3 per cent.

Microscopic Characters.—The groundmass of the rock is holocrystalline and has a fairly uniform grain, averaging 0.05 mm. It is composed of prismoids of orthoclase, micropertite, and anhedral and interstitial quartz, with a sprinkling of dark specks generally square in outline but showing no metallic luster. Plagioclase of the composition Ab₄An₃, quartz, and biotite occur as phenocrysts, forming about 15 per cent of the rock. As no striated feldspar is to be seen in the groundmass, it is probable that all the plagioclase is in the form of phenocrysts, which occur as idiomorphic prismoids, except where they are clustered. The biotite is also in idiomorphic crystals, but the quartz is in rounded or skeleton forms. Magnetite and in a few crystals zircon and hornblende occur as accessories.

Chemical Composition.—The analysis of a perfectly fresh specimen from the summit of the hill gave the following results:

CHEMICAL ANALYSIS OF BIOTITE-SODA GRANITE PORPHYRY OF MARBLE HILL

Constituent	1		2
	Per cent	Molecular ratio	Per cent
SiO ₂	74.23	1.237	73.5
Al ₂ O ₃	13.65	0.134	13.5
Fe ₂ O ₃	0.84	0.005	1.4
FeO.....	1.04	0.014	1.0
MgO.....	0.23	0.006	0.3
CaO.....	0.75	0.013	0.6
Na ₂ O.....	4.87	0.078	3.7
K ₂ O.....	3.96	0.042	5.1
H ₂ O +.....	0.19	..	0.7
H ₂ O -.....			
CO ₂	None		
TiO ₂	0.08	0.001	0.3
ZrO ₂	Present		
P ₂ O ₅	Undetermined		0.1
Cl.....	Trace		
MnO.....	Trace		0.1
	99.84		100.3

1. Biotite-soda granite porphyry, Marble Hill, San Francisco volcanic field, Arizona, H. H. Robinson, analyst. *Specie gravity*, 2.50.

2. Type granite-rhyolite.

¹ ROBINSON, H. H., The San Francisco volcanic field, U. S. Geol. Survey Prof. Paper 76, pp. 105-106, 1913. Research descriptions may be much longer than this. See, for example, A. N. WINCHELL, The gabbroid rocks of Minnesota, *Am. Geologist*, vol. 26, pp. 214-245, 1900.

The rock of Marble Hill is classified as a soda granite porphyry. The ratio of potash to soda is 1:1.9, compared with 1:1 in the type rhyolite.

Mineral Composition.—The mineral composition as calculated from the chemical analysis is as follows:

MODES OF GRANITE PORPHYRY OF MARBLE HILL AND OF TYPE GRANITE-RHYOLITE

Mineral	1	2	3
Quartz.....	29.5	31.4	— 1.9
Orthoclase.....	21.8	28.2	— 6.4
Albite.....	36.1	28.2	+ 7.9
Andesine (Ab ₄ An ₃).....	8.4	} + 3.4
Oligoclase (Ab ₄ An ₁).....	5.0	
Biotite.....	3.1	5.6	} — 2.8
Magnetite.....	0.9	1.2	
Water.....	0.6
	99.8	100.2	+11.3 —11.1

1. Mode of granite porphyry of Marble Hill.

2. Mode of type granite-rhyolite.

3. Departures of mode 1 from mode 2.

INTERPRETATIVE PETROGRAPHY

The most numerous and by far the most interesting problems to settle by laboratory work involve the history of the rock, or some special process or material indicated by the rock.

Problem 12. To interpret textures, mineral grouping and all the microscopic features of a rock in terms of rock history.

Method.—Long experience is needed before some of the features can be correctly understood, but early in the examination of rocks the student should study the criteria tabulated in various parts of this book, and start recording in his notebook his own additions and interpretations of such criteria.

ALLIED METHODS AND APPLICATIONS

Experimental research in petrography and petrology has advanced in recent years largely by laboratory work in fusion and quenching, the methods of investigation of the physical chemist.

The methods of study of rocks are applicable with little change to the study of artificial products—cement, concrete, and the materials of ceramic industries. (See the list of further readings.)

PART II

THE PETROGRAPHY OF IGNEOUS ROCKS

FIELD SCHEDULE FOR EXAMINING IGNEOUS ROCKS

Field schedules for igneous rocks and the economic deposits in them are to be found in several handbooks for field geology. Record should be made of all structures and the lateral and vertical variations, differentiation and assimilation. More elaborate schedules for detailed structural work have been used by Hans Cloos and his students.¹ Some very



FIG. 4.—Flow layers in a syenite, Ely, Minnesota; a primary gneiss.

old masses have been so metamorphosed that the original structure is lost, but others give a great deal of evidence of their form and relations. The features plotted are chiefly:

Flow layers, or layers of different mineral compositions in the intrusive; rows of inclusions, schlieren and schlieren dikes; dip and strike (see Fig. 4). This structure is a primary gneiss of a different sort from that shown in the frontispiece.

Platy and elongated minerals; dip and strike, pitch angle and direction of pitch.

¹ BALK, ROBERT, Primary structures of granite massives, Geol. Soc. America Bull., vol. 36, pp. 679-696, 1925.

Flow lines, or the direction of greatest elongation, commonly marked by the orientation of needle-like minerals (frontispiece), and in some rocks by the greatest elongation of fragments, pebbles and inclusions; pitch angle and direction of pitch (Fig. 5).

Rift and run; dip and strike

Joints and faults; dip and strike, pitch and direction of striation and slickensides.

Dikes and veins; whether aegirine, aplite, pegmatite, or lamprophyre; sequence of dikes and veins; dip and strike; internal structure.

Position of the contact and its nature.

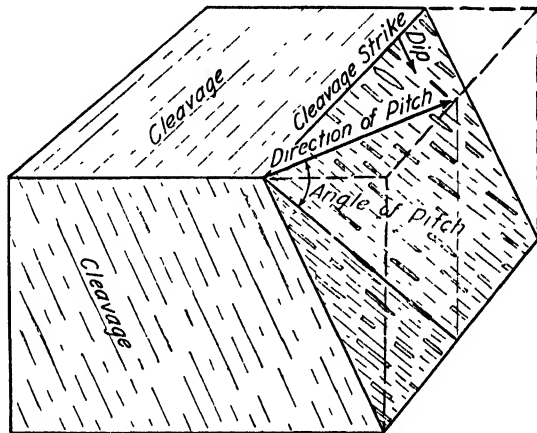


FIG. 5.—Block diagram of flow-line structures to show in three dimensions the features recorded. The dip and strike of the plane structure are correctly recorded by most geologists, but on the cleavage faces or in the plane of the plane structure or flow layers, there are flow lines, oriented needles, striations or stretching. These should also be recorded. The direction of the projection of the flow line on a horizontal surface may be shown by an arrow on the map (Fig. 134). The pitch of the flow lines from the horizontal plane is marked at the head of the arrow.

Country rock, especially near the contact; bedding, dip and strike, drag, and axes of folds; cleavage, dip and strike, and pitch angle and direction of pitch of greatest elongation; dikes and veins in country rock (Figs. 132 and 133).

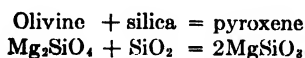
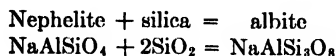
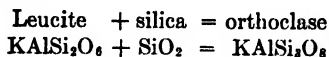
MAIN MINERALS OF IGNEOUS ROCKS

It has been noted (page 4) that in the whole crust of the earth igneous rocks predominate.

The mineral composition of the average igneous rock in per cent is about: feldspar, 59; quartz, 12; amphibole and pyroxene, 17; micas, 4; and accessories, 8.

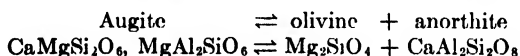
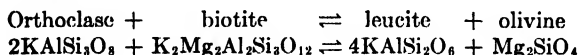
There are certain limits to the mineral occurrences and combinations possible in igneous rocks, the most noteworthy being the fact that certain minerals do not form in association with primary quartz. With very minor exceptions we do not find olivine or feldspathoids with quartz. These minerals react with silica to form other minerals and are not likely

to remain if silica is so abundant as to form quartz. The reactions may be written as follows:



This does not make it impossible, however, for a magma at an early stage to develop olivine and at a later stage, quartz (Fig. 149).

Furthermore, a reversible reaction may make possible a variation in the mineral combinations formed from a single magma, or one in which a very slight change occurred.



It is also noteworthy that calcite and pyrite are questionable as primary igneous minerals, though they are common products of alteration and introduction.

A further set of mineral antipathies is apparently a result of growth at different periods in the evolution of a magma (see page 243).

FORMS, STRUCTURES, AND TEXTURES OF IGNEOUS ROCKS

The terms structure and texture are loosely used and not sharply distinguishable. In general, structure in a rock refers to the larger features of form, occurrence, jointing, etc., whereas texture refers to the relation of grain to grain in the rock. Where grains are small, however, a feature as small as an amygdale may be referred to as a structure.

The confusion is increased by the fact that internally a crystal may show structures, such as zoning, wavy extinction, twinning and secondary growth, for which we have no other term than structures. Structure should, therefore, always refer specifically to a rock, or to a mineral grain. (See the list of further readings.)

STRUCTURES

Extrusive igneous rocks are those that are poured out as lava flows at the surface of the earth. If eruption occurs explosively, the rocks are *fragmental* and may be called *pyroclastic*.

Intrusive rocks are those that solidify from magma underground (see the tabulated criteria for distinguishing extrusive forms from intrusive sheets, page 32).

Dr. Hans Cloos¹ has recently proposed a series of terms for the forms and structures of intrusives, based on the observed features and avoiding assumptions as to the contacts not exposed. The exposed contacts may show the mass to be *periclinal*, *synclinal*, *vertical*, or *horizontal*. He also suggests terms to show the relation to tectonics; the relations of internal and external structures to each other and to the contact (page 198).

CRITERIA FOR DISTINGUISHING SURFACE FLOWS AND INTRUSIVE SHEETS

Extrusive

1. Upper surface likely to be scoriaceous, pumiceous or vesicular. Amygdaloids.
2. Flow structures (distinguish from flow structures in metamorphic rocks), ropy, twisted masses.
3. Glassy textures. Crystallites: Trichites, belonites, globulites, spherulites, lithophysae (stone bubbles), perlitic cracks, vitrophys.
4. Phenocrysts in some thick flows. Many corroded.
5. When extrusive sheets are overlaid by sedimentaries, the basal part of the sediment is likely to contain fragments of the flow.
6. Base of an overlying sediment not metamorphosed at the contact. Little metamorphism even below.
7. No sediment or roof rock included in upper part of the igneous rock.
8. No apophyses of igneous rock in overlying rock.
9. Top of igneous rock an erosion surface.
10. Many with ash and breccia.
11. All have chilled margins.
12. Few have graphic groundmass.

Intrusive

1. No extensive scoria or amygdaloid.
2. Flow structures rare.
3. Glassy only in small sheets or along contacts. No crystallites.
4. Phenocrysts more likely and larger. Less corroded.
5. Overlying clastic rocks contain no fragments of the igneous rock.
6. Beds above as well as below have been affected by heat.
7. Contact breccia in upper part of the igneous rock.
8. Overlying sediments may be cut by stringers and dikes—offshoots of the main mass.
9. The intrusive sheet itself may cross sedimentary beds.
10. Not usually associated with ash.
11. May be coarse to the contact. Many are diabasic.
12. Graphic groundmass if they have quartz and acid feldspar.

Daly makes a valuable distinction between *injected* bodies, which come to occupy their chambers through channels more or less restricted



FIG. 6.—Section through Ear Mountain, Alaska. (After Knopf.) A stock increasing in size with depth.

connecting the chamber with a deeper reservoir or source, and *subjacent* bodies, which seem to be parts of the deeper reservoir itself and have roots rather than a floor.

¹ CLOOS, HANS, *Fennia*, vol. 50, No. 2, 1928.

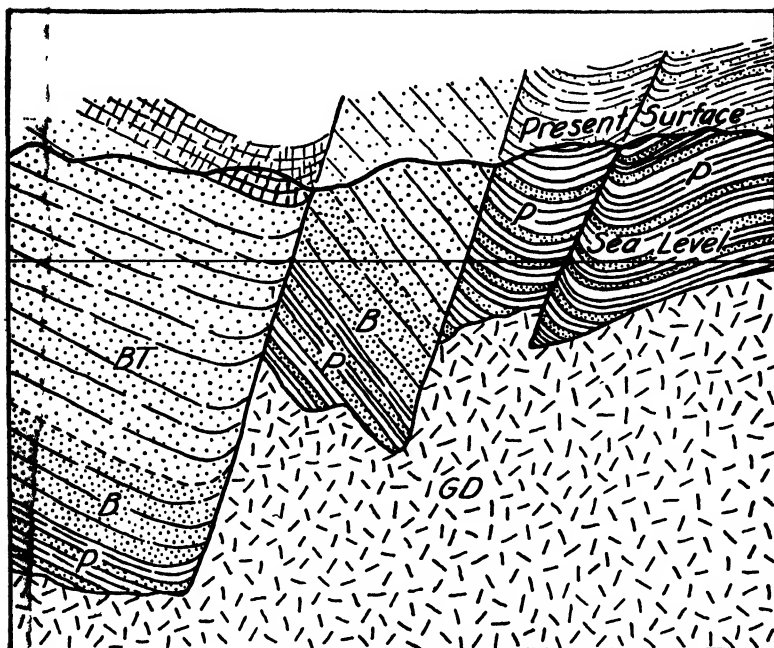


FIG. 7.—The upper contact of an intrusive in Idaho. (After Sampson, *Idaho Bur. Mines and Geology, Pamphlet 31.*) The blocky nature of the contact strongly suggests that fragments of the roof have been stoped in.

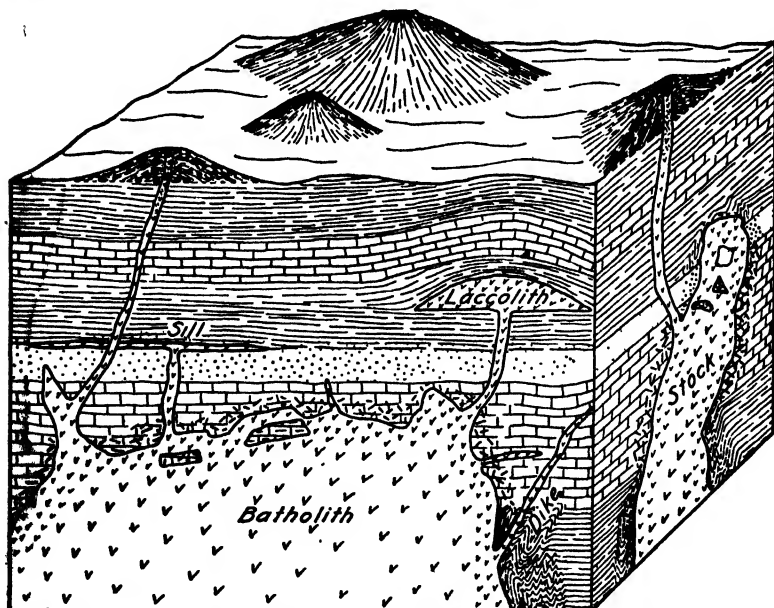


FIG. 8.—Block diagram of an igneous area showing the supposed relations between the several forms of igneous rocks. (Modified from Butler.)

Probably the largest bodies of igneous rocks in the earth's crust are *batholiths*, such vast bodies that even the great depth reached by all erosion since Laurentian time fails to show the nature of the floor or roots. Such masses are *plutonic* or *abyssal*. Most batholiths lie in *orogenic belts* elongated with the mountain range structures. Most of them were

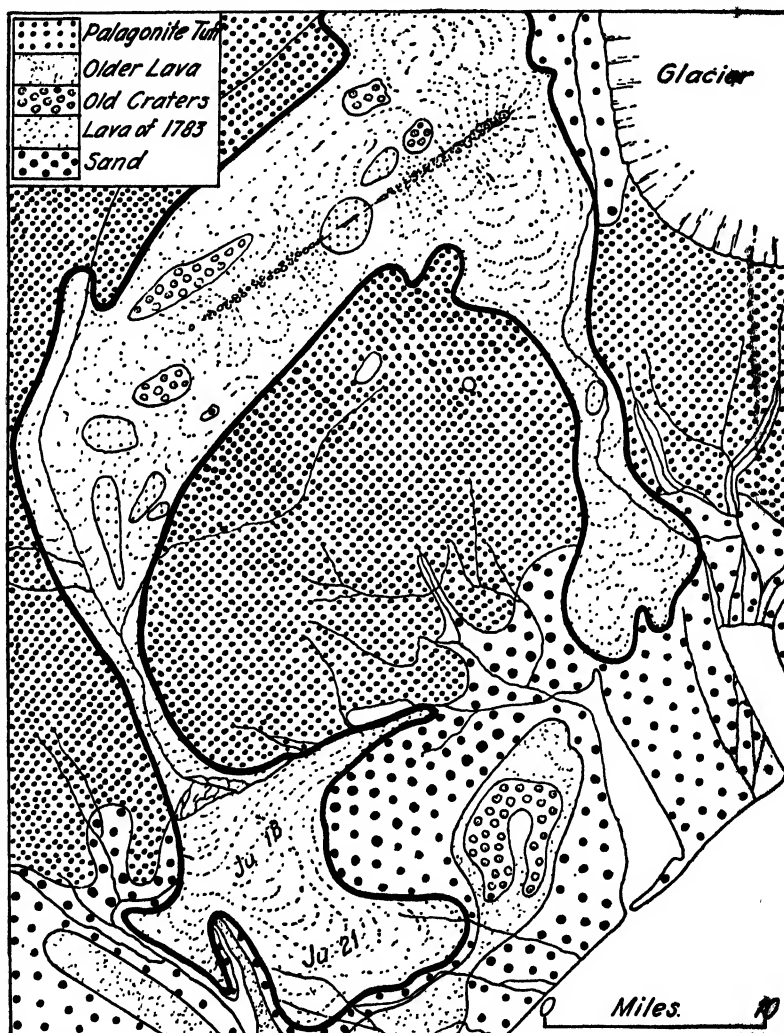


FIG. 9.—Lava flow of the plateau type or fissure eruption type, Laki, Iceland, 1783. (After Helland.)

intruded soon after the folding and deformation of their walls and roof, but in the formation of some, invasion and mashing alternated. Harker noted the distinction between "mountain" and "plateau" batholiths. The batholiths transgress earlier structures in their roofs and walls,

but many induce a schistosity in the walls parallel to the contact. The walls are steep, and the contacts range from smooth to blocky (Figs. 6, 126, 135, and 144). The magna may rise to within 2,500 feet of the surface, but no doubt many batholiths solidified at great depths. Some large masses intruded schistose or bedded rocks in very characteristic layers (see Fig. 223 and frontispiece), and the term "stromatolith" may be used for the mixed, generally concordant mass. *Stocks* seems to

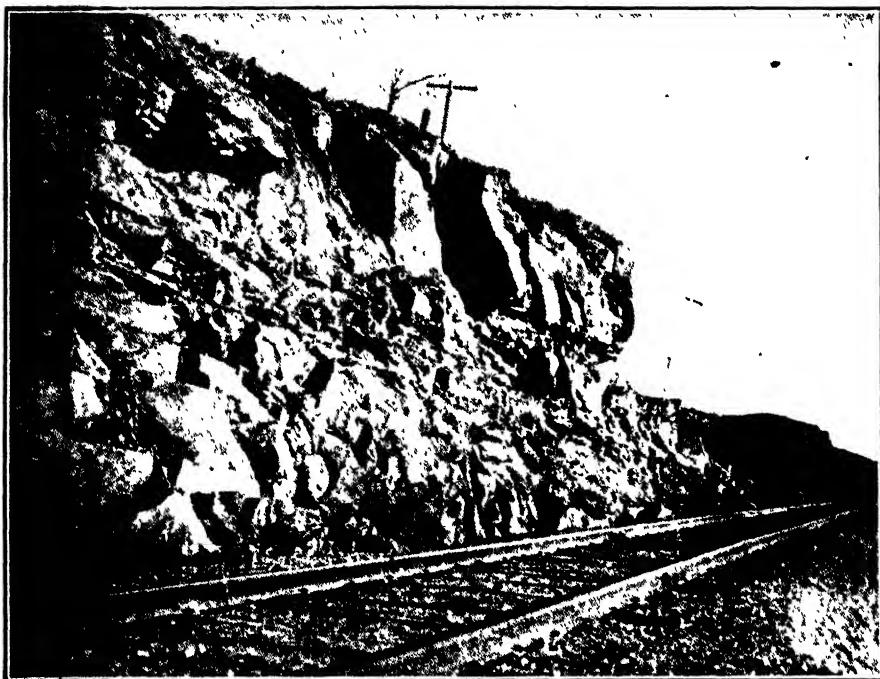


FIG. 10.—Two lava flows exposed by erosion at Duluth, Minnesota. The dense hard overlying flow rests on the soft more-jointed amygdaloidal upper zone of the lower flow.

be essentially small batholiths not more than about 20 square miles in area, or cupolas rising above the general level of the batholith roof.

The large areas of *surface flows* of lava (Fig. 9) and their fragmental equivalents in *surface beds*, are nearly equal to the exposed areas of batholiths, but even in the great lava fields, the flows must have a relatively small thickness (Fig. 8).

Intermediate in position (*hypabyssal*) are the *dikes* (Fig. 11 and pages 135 and 190), *plugs* or *necks*, *sills*, and *laccoliths*, forms familiar to every student in geology. Laccoliths with roofs faulted up are *bysmaliths*. Fewer but of enormous size are the centrally sunken lopoliths (Fig. 14). The *ethmolith* is a funnel-shaped intrusive; the *sphenolith* is wedge shaped. Small dikes and stringers related to a larger mass and near to its edge are *apophyses* (Fig. 12); some apophyses in bands along the cleav-

age planes of a slate have been described as *ribbon injections*. Larger intrusions with schist walls may intimately soak into the leaves of the schist forming a *lit-par-lit* injection (frontispiece). Intrusive masses with less regularity of form are called *chonoliths*. A surprising number of intrusives are so irregular that no name has been given them but the miners' name "*blowout*."

Most of these injected bodies may show different degrees of complexity, distinguished as *simple*, *multiple*, and *composite*—multiple if there were several injections of the same kind of magma; composite, if there were several of different kinds of magma.



FIG. 11.—A black dike in granite, Saganaga Lake, Minnesota.

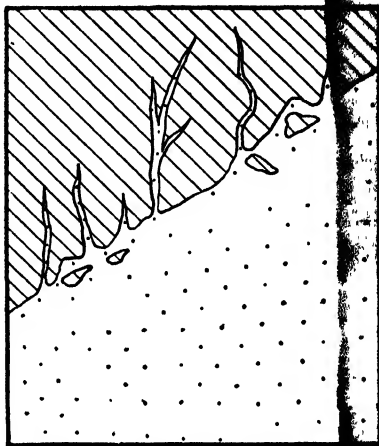


FIG. 12. Contact of gabbro at Duluth, Minnesota, and its roots of older basalt. Apophyses of the intrusive rock (stippled) extend into the older rock (diagonal lines) showing clearly the once fluid nature of the gabbro material. Area of exposure about ten feet square.

Xenoliths are rock fragments and *xenocrysts* are mineral fragments included in a magma. *Cognate xenoliths* are fragments of an early chilled border of the same magma.

Orbicular structures occur in granitoid rocks and consist of concentric, many of them radial, aggregates, the largest a few inches in diameter. Some have a xenolith as a core or nucleus (Fig. 25).

Once magma is in a place of rest, a number of geologic processes produce further structures in it. Bunchy segregations, cells or vesicles, and miarolitic cavities develop during cooling. Later the shrinkage of the solidified rock causes cracks and joints, platy, coordinate (Fig. 177) columnar (Fig. 16), spheroidal, perlitic, and others; and the several geologic processes produce faults, folds, slickensides, and other structures that are common in both igneous and other formations. Possibly most noteworthy is the columnar jointing of tabular masses. If the sheet is

irregular or has a connection with deep masses, the columnar structure may be irregular too. A tabular or sheetlike jointing is commonly developed parallel to an exposed surface of a rock and may be attributed to changes of temperature affecting the surface. Spheroidal structures commonly develop in a granitoid rock by weathering.

Some structures in addition to the almost universal joints are fairly common in igneous rocks. They may be distinguished as flow layers and flow lines. The flow layers are alternations of mineralogically unlike layers or lenses (Fig. 4). The division may be sharp or gradual, but the textures differ very little, and the minerals interlock across the contact.

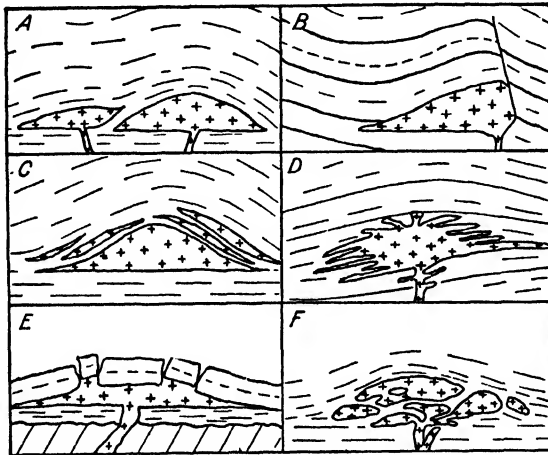


FIG. 13.—Laccoliths and related forms. A, Henry Mts., Utah. (After Gilbert.) B, Mt. Marcellina, Colorado. (After Cross.) C, Judith Mts., Montana. (After Pirsson.) D, La Plata Mts., Colorado. (After Holmes.) E, Black Hills, South Dakota. (After J. D. Irving.) F, El Late Mts., Colorado. (After Cross.)

Many of the layers are lenticular, and they differ in color and in the proportion of vesicles of glass, etc. Possibly some of these differences are results of an original heterogeneity in the magma. Fluxion structure is a parallelism of the grains that are platy, needle-like or elongated in any way (Figs. 17, 49, and frontispiece). Most of the rocks that have primary flow layers have fluxion structures in some of the layers parallel to their length. If the mineral elongation is greatest in one direction, the position of this greatest elongation may be described as a linear foliation, or flow line (Fig. 5). Finally, nearly all the rocks showing these structures develop a jointing parallel to the flow layers and independent of the joints due to weathering. The platy parting in rhyolite intrusions is especially notable.

Rocks showing flow layers and other plane structures like these are commonly metamorphic, but if careful examination shows no granulation or recrystallization such rocks may be igneous. These rocks are called "primary gneisses" by many petrographers to distinguish them from

gneisses that are of metamorphic or injection origin. The qualifying adjectives *trachitoid* or *gneissoid* for coarse rocks and *trachitic* for felsites are here regarded as preferable, the name gneiss being largely employed

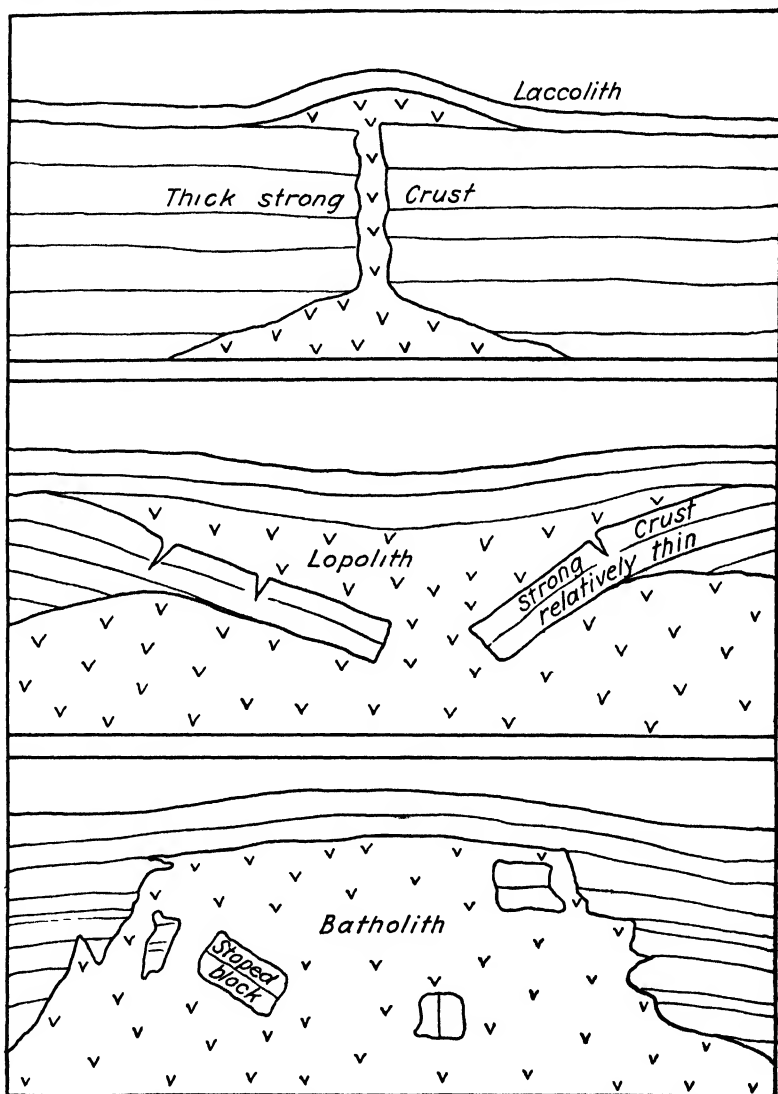


FIG. 14.—Diagram suggesting the probable nature of a lopolith, intermediate between laccoliths and batholiths.

for metamorphic rocks. The term “foliated” is also applicable but is mostly used for rocks banded by metamorphism (see later section).

It is commonly stated that igneous rocks are “massive” in structure if they show none of these special structures. Nevertheless, although

many rocks both intrusive and extrusive seem massive on superficial examination, a detailed study shows a parallelism of the elongated feldspars. This is attributed to some movement in the magma during

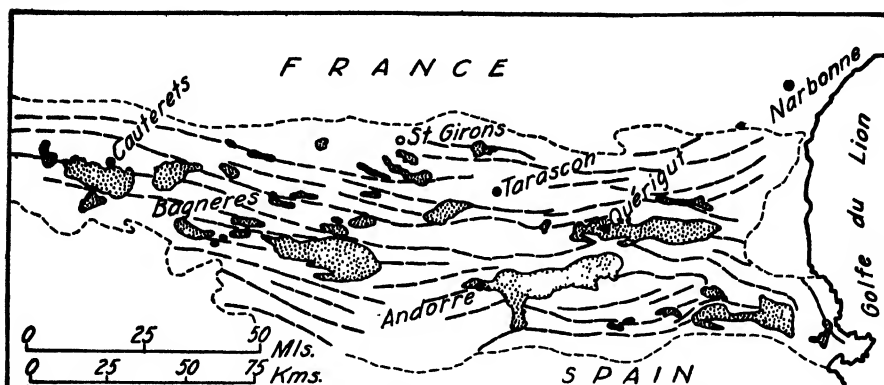


FIG. 15.—Map of some batholiths and stocks (dotted) in the Pyrenees. (After Daly.) The igneous masses are clearly elongated in the direction of the orogenic axes, the trends of which are shown by the broken lines.

crystallization. It is recognized by quarry workers as a direction of easy splitting. In trimming out blocks from a quarry the direction of easiest splitting is the "rift," and the easier of the two at right angles to this is the "run" or "grain" (see Fig. 26).

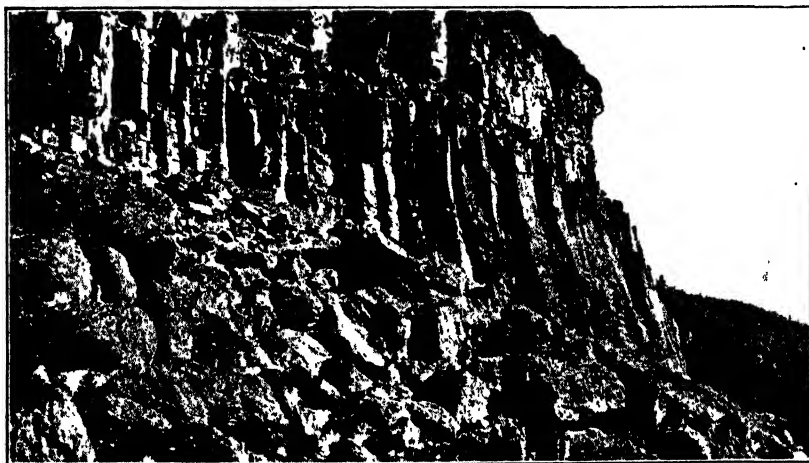


FIG. 16.—Columnar jointing in a lava, Obsidian Cliff, Yellowstone Park. (Photo by Balk.)

The form, size and condition of solidification of an igneous mass are largely responsible for the textures developed in the resultant rock. The composition of the magma and modifications in the several factors affecting texture will be taken up later.

TEXTURES

Textures of igneous rocks may be considered as involving three features; first, crystallinity, the proportion of crystals to glassy material; second, granularity, the size of the crystals; and third, fabric, the arrangement and relations of the crystals. The textural terms in common use do not fully cover all these factors.

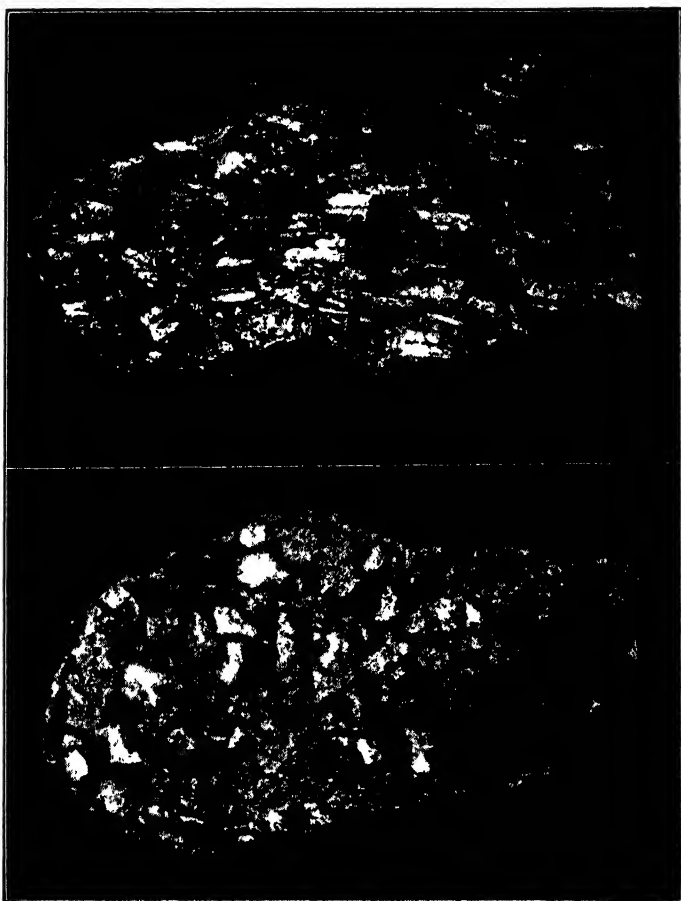


FIG. 17.—Side and top views of a rock showing fluxion structure, Duluth, Minnesota. White plagioclase plates are surrounded by ferromagnesian minerals. About natural size.

A rock may be *crystalline*, or *hyaline* (glassy), or *hypocrystalline* (partly crystalline). *Cryptocrystalline* rocks are so fine grained that their crystalline character is recognized only by a confused aggregate polarization.

The size of grain in igneous rocks is so variable that it is difficult to set satisfactory limits. Rocks in which a notable proportion consists of

glass or grains too small to be distinguished in hand specimens are said to be *aphanitic* (aphanites), and those in which all grains are visible except a very small proportion of minute crystals are said to be *phaneritic* (phanerites). A phanerite is said to be *fine grained* if the main grains are less than 1 mm. in diameter and *coarse grained* if they are more than 5 mm. in diameter.

As to granularity many igneous masses are *equigranular*. Inequigranular textures may be *seriate* or *hiatal*. If there is a hiatus between large and small grains, the finer material is *groundmass*, and the large crystals are *phenocrysts*; the texture is *porphyritic*. The groundmass is also referred to as a base, or basis, or matrix.

Very minute crystals in a glass are *crystallites*, or if too fine, *dust*. Crystallites may be classed as *belonites* if rodlike, *globulites* if rounded, and *trichites* if hairlike. Radial groups are *spherulites*, or in a basaltic rock *variolites* (see Fig. 91). *Lithophysae* are hollow and concentric spherulites. Crystals that crowd out their neighbors into a peripheral pattern make the rock texture "ocellar."

The forms of crystals are *euhedral* if the outlines are straight and regular (automorphic and idiomorphic are terms of similar meaning). Crystals are *subhedral* if they closely approach straight regular outlines. They are *anhedral* if irregular, having been interfered with by the growth of adjoining crystals (xenomorphic and allotriomorphic are terms of similar meaning).

Most phenocrysts are euhedral or nearly so. Phenocrysts that have been partly dissolved, rounded or embayed are called *corroded*. Special terms often applied to the euhedral forms—tabular, platy, acicular, octahedral, equidimensional, etc.—are well known. A mineral that occurs in both phenocrysts and groundmass may be said to occur in *two generations*.

Common internal structures in crystals are *twinning*, *zonal growth* (see page 89), and an intergrowth with some other substance or a collection of inclusions.

The textural series in each of the main igneous rock clans is approximately the same, only a few special textures being restricted to a particular clan. Furthermore, in each clan there is found to be a general relation between texture and occurrence, so that a particular texture suggests at once the probable conditions of origin and the form. The relation, however, is so uncertain that the laboratory petrographer should not commonly assume that the form is thus determined. The standard textural terms of general application are here tabulated, and special terms applicable only to certain clans are given later. The order of arrangement is based on approximate position in the earth at the time of formation, not with the idea of emphasizing the surface beds, which are named first.

TABLE II.—STANDARD TEXTURES AND OCCURRENCES OF IGNEOUS ROCKS

	Textural Terms	Suggested Occurrence
Aphanitic	Fragmental, tuffaceous, brecciated	Surface beds
	Glassy	
	Felsitic = aphanitic, but crystalline, without phenocrysts	{ Usually surface flows; effusive; extrusive; volcanic
	Aphyric = aphanitic, usually crystalline, without phenocrysts	
	Trachitic = aphanitic with flow structure	
Phaneritic	Porphyritic = phyrlic { vitrophyric, with glassy groundmass felsophyrlic, with crystalline groundmass	{ Usually intrusive at moderate depths, hypabyssal
	Prominent vs. predominant phenocrysts	
	Granitoid: Equigranular	{ Usually subjacent or plutonic masses; abyssal; deep-seated intrusives
	Porphyritic	
	Aplitic = sugary textured	
	Pegmatitic = very coarse, variable	
	Gneissoid = (or trachitoid) = granitoid in size of grain but with flow structure	

Several of these terms are defined in the preceding paragraphs, and most of them are familiar even to elementary students in geology. Several are used in compound rock names, e.g., grano-, vitro-, and phyro-. Hyalo- is equivalent to vitro-.

Most of them may also be used in names of rocks, if a qualifying adjective shows the composition, as rhyolitic vitrophyr, syenite aplite, and dacite porphyry. Some of the textures (such as trachitic) would have a compositional meaning if used as nouns. In general, it seems best to leave the textural term in a minor position in naming rocks, on account of the more or less accidental and local development of textures. The adjective granitic is sometimes used as a textural term but is best restricted to indicate composition; granitoid is a better textural term.

Granitoid normally indicates a phanerite of medium grain, approximately equigranular. The presence of some euhedral grains may be shown by the prefix hypautomorphic. "Monzonitic" has been suggested as a term for hypautomorphic granitoid (Fig. 59), but this also is open to criticism as indicating composition rather than texture.

Special Textural Terms.—*Graphic*, micrographic, micropegmatitic, and granophyric refer to regular intergrowths of two minerals, commonly quartz and acid feldspar, with a pattern resembling cuneiform characters. Myrmekite is a more wormy intergrowth due to reaction.

Ophitic and *diabasic* refer to an intergrowth of plagioclase laths and anhedral pyroxene. The feldspar is euhedral, and pyroxene, commonly in larger grains, fills the interstices or surrounds the feldspar. If there is any distinction between the two terms, it is probably best to use diabasic when the feldspar is in relatively large and abundant grains.

Poikilitic refers to the inclusion of many large grains of one mineral in still larger grains of another (see Fig. 80b). *Ophitic* describes a special combination of augite enclosing plagioclase.

Mosaic and *granulitic* refer to equigranular rocks with grains smoothly in contact with their neighbors and about the size of granulated sugar. The texture is common in aplites and in metamorphic rocks.

Sutured refers to the very irregular interlocking contacts of grains in approximately equigranular rocks.

Pilitic means feltlike, resembling a mass of slender fibers in an irregularly interlocking network.

Intersertal describes needles of plagioclase with a variety of materials filling the interspaces.

Glomeroporphyritic refers to a porphyry in which the phenocrysts occur largely in clusters, somewhat separated from other clusters (Fig. 62).

Eutaxitic refers to a banding in surface lavas, commonly the result of cohesion of pasty fragments of lava or tuff.

Vesicular and *cellular* refer to bubble-like cavities, most of which are found in surface lavas.

Amygdaloidal is a term to describe the secondary filling of vesicles (Fig. 70).

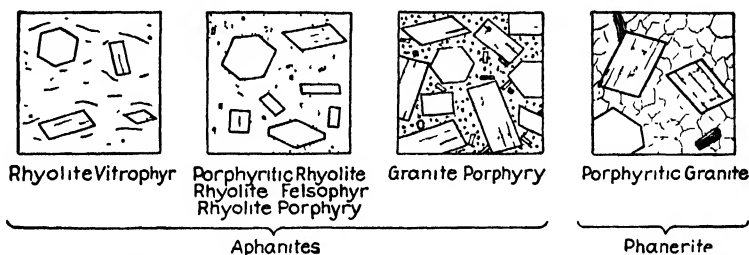


FIG. 18.—Diagrams of the textural differences between the several porphyritic rocks of the granite-rhyolite clan (see also Figs. 22 and 23).

Combination may occur, such as hyalopilitic, or a porphyry with a mosaic groundmass.

It should be noted that porphyritic rocks in any clan may be of several kinds. (1) The common intrusive or extrusive porphyry has a fair sprinkling of phenocrysts in an aphanitic groundmass. If the mass is a flow, it is named from the composition, with the prefix porphyritic. If an intrusive, it is named a porphyry, with a prefix to show the composition. The nature of the groundmass is the basis of a further distinction between vitrophyrs and felsophyrs. (2) If phenocrysts are predominant, say more than 50 per cent, the abundance of phenocrysts characterizes a granite porphyry, for example, in distinction from a rhyolite porphyry with less abundant phenocrysts but with the same composition. Usually a rock with abundant phenocrysts has a coarser groundmass than one with few. (3) Finally, a granite with phaneritic groundmass may have phenocrysts in the sense that a few grains are relatively large and euhedral. Such a rock is a porphyritic granite and should not be confused with the porphyries that are aphanitic (see Figs. 18 and 23).

ORDER OF CRYSTALLIZATION

Using the criteria indicated by Fig. 19, Rosenbusch studied a large number of igneous rocks and estimated that the normal order of crystallization was (1) the group of ores, (2) the ferromagnesian minerals, (3) the feldspars and feldspathoids, (4) quartz. It is not to be assumed that one mineral finishes its growth before the others begin, but rather that there is some irregular overlapping in time. There are some exceptions, such as diabase, and recently the whole subject has been opened up by a

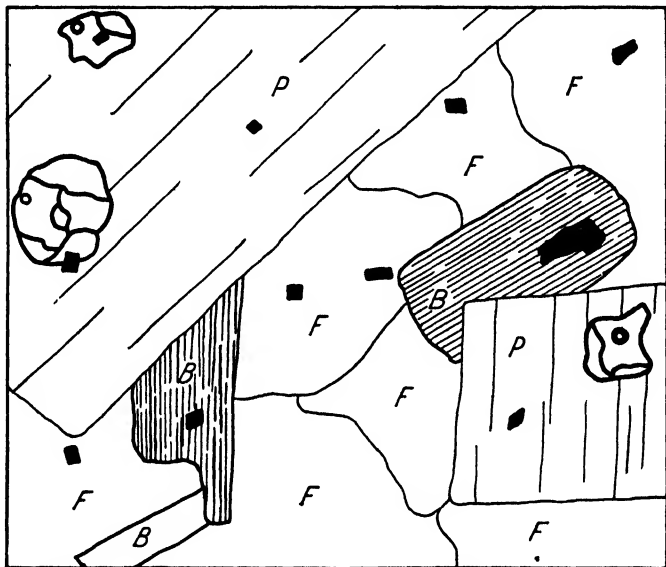


FIG. 19.—Sketch showing the common textural relations used in estimating the order of formation of igneous minerals.

criticism of the criteria themselves. Euhedral crystals included in larger grains may be introduced, but most petrographers still consider euhedral inclusions in igneous minerals as earlier than the host (see the criteria, page 441).

DISTINGUISHABLE STAGES OF MAGMATIC ACTION

A considerable terminology has recently grown up relating to stages of magmatic action. Several are elaborated in later pages, especially from pages 242 to 246.

OUTLINE OF STAGES

Formation and accumulation of magma.

Intrusion; possibly stoping; assimilation of wall rock and fragments; schlieren.

Streaming of gases; if pressure is reduced by intrusion there are early emanations; early pegmatites; the emanations cause exomorphic contact effects, and internally also some differentiation in the magma.

Partly miscible separation.

Normal crystallization, orthotectic stage; accompanied by more or less diffusion, convection and segregation by crystal settling or floating.

A *sequence* of crystallization. Some late simultaneous growths. A "second generation" of crystals in groundmass; zoning; corrosion; resorption. Continued **stopping**, assimilation and streaming.

Pegmatitic stage; high vapor pressure causing emanations above critical temperature; a series of reactions of crystals and magma; deuteric effects; paulopost effects; pneumotectic effects; graphic, myrmekitic and syntectic minerals.

Segregation by interference of crystals in the moving "mush."

Filter pressing, or straining out of mother liquor; pegmatites and aplites.

Late emanations; amygdules; replacements in pegmatites.

Mineral transformations and unmixing of solid solutions; reactions between adjacent solid crystals.

Later than these come the hydrothermal alterations, which are not usually classed as magmatic alterations, though the hot waters may be of magmatic origin.

DEVELOPMENT OF THE CLASSIFICATION OF IGNEOUS ROCKS

The Rosenbusch system, which may be said to emphasize occurrence above texture and to be qualitative rather than quantitative, has dominated petrographic usage for many years. About the beginning of this century a group of petrographers in the United States decided that the next advance in petrography should make the classification more exact and quantitative (see Problem 8, page 20). In their attempt to do this they selected as a basis the quantitative chemical analyses of rocks and subordinated the actual minerals as well as the texture. In Germany Osann, Niggli, and Hommell have similarly emphasized quantitative chemical analyses. This movement has resulted in real progress, but the tendency to ignore the actual minerals of the rock has aroused so much protest, and the expense of analyses is so great that the new systems have not replaced older ones to any great extent. Quantitative chemical classifications have their place in advanced and regional petrology but are not very serviceable in direct practical applications of the subject.

The quantitative idea was taken up in other proposed classifications based on actual minerals. Several were well planned but used the common rock names with such radically revised definitions that the schemes have not been adopted. More recently a number of other petrographers have proposed classifications of igneous rocks in which the tendencies of recent years are clearly shown (see the list of further readings). The chief features of these schemes are:

1. Most of them are more exact and quantitative than older schemes.
2. Most of them subordinate detailed chemical features and textures to mineral composition; texture being an accidental feature related to rate of cooling and being variable in short distances.

3. Occurrence is left for another system of classification—the geologic system.

4. The minerals are *actual* rather than standard, calculated or possible minerals or pure molecules. Magmas of similar composition may crystallize to different minerals (see page 31).

5. The main groups are distinguished by minerals that differ notably in either *alkali* content or *silica* content. This is in sharp contrast to

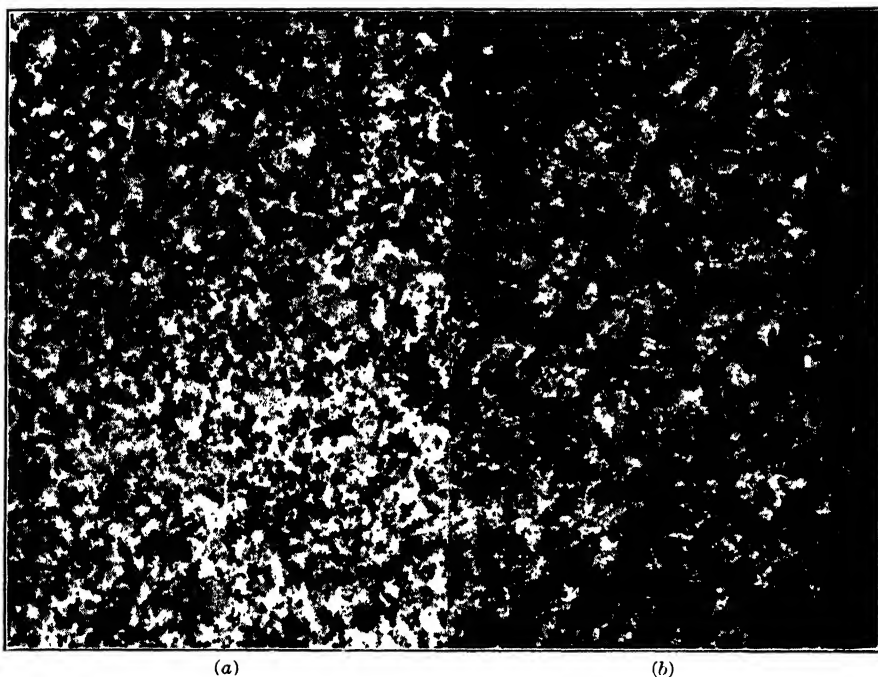


FIG. 20.—Granitoid rocks of different color tones. (a) Granite, a leucocratic rock, Giants Range, Minnesota. (b) Gabbro, a melanocratic rock, Salem Neck, Massachusetts. Natural size.

some older schemes based on the differences between hornblende and pyroxene, or even on minute features such as the difference between diallage and augite.

High silica is marked by quartz; low silica by feldspathoids, by orthosilicates like olivine and by oxides. The “saturation” is expressed sufficiently well by mineralogic prefixes.¹

6. The distinction between light and dark minerals can be made systematic and quantitative only by listing the minerals. The terms “salic” and “femic” refer to calculated *molecules*, not actual minerals, and should be used only in connection with chemical composition. The terms *felsic* and *mafic* have come into fairly wide use; they have mne-

¹ SHAND, S. J., On saturated and unsaturated igneous rocks, *Geol. Mag.*, vol. 50, p. 508, 1913.

monic derivation: (fe) for feldspar, (l) for lenads or feldspathoids, (s) for silica, (ma) for magnesia, and (f) for iron. Felsic minerals include feldspars, feldspathoids and quartz; presumably also corundum, zircon, and muscovite. Mafic minerals include "ferromagnesian minerals of all kinds"; amphiboles, pyroxenes, biotite, garnet, tourmaline, spinel, olivine, iron oxides, and ilmenite; presumably also apatite, titanite, perovskite, fluorite, and metals and sulphides. The following terms may also be found useful.

Adjective	Prefix	Significance
Leucocratic . . .	Leuco-	Light colored
Mesocratic	Meso-	Medium colored
Melanocratic	Mela- or melano-	Dark colored

7. The feldspars are used in classification very differently by different petrographers, but the tendency is clearly to distinguish more than two groups. Rosenbusch early distinguished *orthoclase* from *plagioclase* and later added a rock group in which the two are nearly equal. Lindgren called attention to the abundance of granodiorites (page 87) and to accommodate them divided the quartz-feldspar rocks as follows:

More than two-thirds of the feldspar is orthoclase	Granite
From one-third to two-thirds of the feldspar is orthoclase	Quartz monzonite
From one-third to $13\frac{1}{3}$ per cent of the feldspar is orthoclase	Granodiorite
Less than $13\frac{1}{3}$ per cent of the feldspar is orthoclase	Quartz diorite

Other men distinguish "alkalic" and "soda-lime" feldspars; some even dividing the soda-lime series, calling the more calcic end "lime-soda" feldspars. Finally, some men are distinguishing silicic, medium, and basic feldspars. As the silicic feldspars are exactly the same as the alkalic, it is possible to subdivide the silicic group into sodic and potassic.

Instead of the terms alkalic and silicic, here used to characterize certain feldspars and rocks some men have used "alkaline" and "acid." These terms are used by chemists for materials of specific and opposite reactions. It is confusing to anyone trained in chemistry to find a rock described as acid and alkaline at the same time, but there are many minerals and rocks that have abundant silica associated with more than average quantity of soda and potash. These are properly silicic and alkalic too.

Attention must here be called to the fact that *no single* series of isomorphous feldspars is a satisfactory basis for a complete classification, for it can not include those associated feldspars characteristic of monzonites. Neither is the series just quoted from Lindgren wholly satisfactory, for it says nothing of the nature of the plagioclase. Both isomorphous and more mechanically associated feldspars should be taken into account.

Rock feldspars are so complex, however, that even a double series, one isomorphous and one mechanical, will hardly yield a solution of all difficulties; for the feldspars that are zoned with different compositions in successive zones are not given separate consideration; average compositions are usually estimated for such grains.

Figure 21 probably illustrates the progressive usage of feldspars at present. In this scheme the varieties are named according to their ratio of silica and base; but it is recognized that in medium and more basic feldspars there is little potash, whereas in silicic feldspars there may be orthoclase, microcline, albite, and various intermediate, isomorphous, and intergrown feldspars. Attention may also be called to the fact that

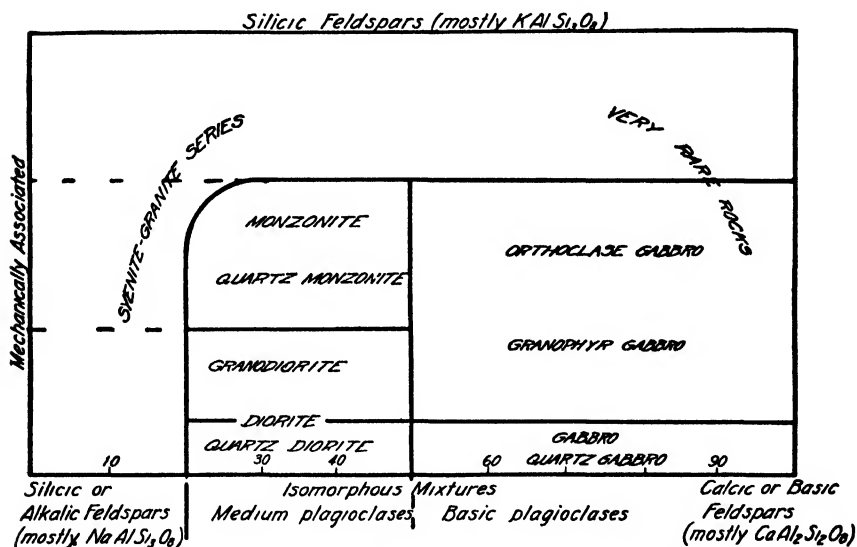


FIG. 21.—Current methods of using feldspars in the classification of igneous rocks.

all feldspars with index less than balsam fall in the silicic group; also that the line between medium and basic feldspars is at 50 per cent anorthite, as is customary.

8. It has seemed to most petrographers that the use of feldspar and feldspathoid in primary classification placed sufficient emphasis on alkalis without further dividing large groups on that basis. Shand notes, however, that whereas the ratio of alkali to silica is 1:6 in feldspar and mica, the ratio is greater in acmite, arfvedsonite and the feldspathoids. In this sense a rock is alkalic if it has minerals in which the ratio of alkali to silica is greater than 1:6. We may have alkalic granites as truly as alkalic syenites, quite regardless of the abundance of feldspar, though usually the several signs of abundant alkali are associated. The prefix "alkalic" serves to distinguish such varieties (see the list of further readings).

Mineral	Alkali	SiO ₂	Al ₂ O ₃
Feldspar.....	1	6	1
Mica.....	1	6	3
Leucite.....	1	4	1
Acmite.....	1	4	0
Nephelite.....	1	2	1

Before proceeding to tabulate a classification on these principles, it should be noted that certain other features are much to be desired if at all feasible: (1) Natural groups should be together; (2) old and well-established names are good, and the definitions should not be changed; (3) textures are taken up later in detail as varieties coordinate with mineralogic varieties in each clan; (4) to use age as a basis of classification, as has been done by certain German and French petrographers, is certainly not to be recommended; possibly the qualifiers "cenotypal" and "paleotypal" may have real meaning if prefixed to some rock names. The devitrified glasses and propylites serve to illustrate paleotypal rocks.

NOMENCLATURE OF IGNEOUS ROCKS

The custom of systematists in petrography has long been to select a new name for each new species and even for varieties, basing it on some property or the name of a locality. The suffix -ite is commonly added to complete the name.¹

The number of names for igneous rocks invented in recent years is so great that it is practically impossible to remember their meanings, and it may be added that it is not worth while to remember the meanings of most of them. H. S. Washington² says:

It seems better to revert to the simple method of employing certain fundamental rock names such as granite, rhyolite, syenite, trachite, diorite, andesite, gabbro, and basalt, which by long and general use and connotation imply the broadest and most fundamental characters of mode and texture, and which are to be modified by qualifiers to denote more specific mineral or textural features or peculiarities . . . The total number of fundamental names would never be very large, probably not much over one hundred or so . . . and their meanings being broad and simple could be remembered with comparative ease.

If this point of view can be agreed upon, it remains to give accurate definitions to the fundamental names. Up to the present time they have been only loosely defined, and many petrographers feel it necessary to define their terms every time they use a rock name. Certainly every

¹ The same ending is used for mineral names and a suggestion was once made that rock names be made by using -yte as an ending. This still appears in the terms trachyte and trachytic, but not in other common names.

² WASHINGTON, H. S., Petrology of the Hawaiian Islands, Am. Jour. Sci., 5th ser., vol. 5, pp. 472-473, 1923.

petrographer when giving a rock an exact name, or a name on which some importance is placed, should add enough qualifying or descriptive terms to make his meaning perfectly clear.

Petrographers are almost everywhere hoping for a standard nomenclature, but they have been slow to accept any proposed standard except their own individual opinions. At one time the petrographers of the U. S. Geologic Survey agreed on certain practices for the names used in folios of the Survey, but these were not published nor accepted by all. British geologists have had a Committee on Nomenclature, which has made good suggestions and probably some real progress. They base variety names on the well-known names of common rocks. They regard mineralogic qualifiers as greatly to be preferred to locality names. If the varieties found at some locality are peculiar in texture or accessory minerals, they suggest giving the rock the common name and following that in parenthesis with the explanatory notation of the type, for example, nephelite-syenite (Foya type) rather than foyaite. They use a hyphen to connect a mineral to a rock name, if the combination is for a species, but change the mineral name into adjectival form for minor varieties, for example, biotitic quartz-diorite. Gradations may be indicated by hyphenating the two rock names to which the specimen is intermediate, for example, gabbro-diorite, or monzonite-syenite.

The table here presented for granitoid rocks divides them into families. It has seemed desirable further to enclose with a heavy line certain related families, forming what may be called a *clan*, for facility in description.

Notes on the Classification. *Note 1.*—This classification is not presented as a definitely progressive suggestion or as a reform in nomenclature of petrography, much as the names may need revision. It is an attempt to set quantitative limits, even if unavoidably broad, for the names commonly in use, and to set them in agreement with the common usage. More restricted definition can then be given for subdivisions in each clan. Textural and mineralogic divisions may be called rock *species*, but minor exceptional rocks are *varieties*.

Note 2.—The alkali content of a rock is to be indicated by a prefix or qualifying adjective. Any attempt to distinguish three degrees of alkalinity gives undue emphasis to a very rare group of "peralkalic" rocks.

Note 3.—The proportion of light to dark minerals may be worthy of more emphasis than is given in the table, where rocks are divided into two groups, one more and one less than 90 per cent of mafic minerals. For example, it has been suggested that a primary division of igneous rocks into four classes may be made on the proportion of felsic and mafic minerals (see Fig. 20), and considering the second group, with 5 to 50 per cent mafic minerals, as medium for many rock families; the prefix "leuco-" could be used for lighter rocks and "mela-" for darker rocks. This seems to disregard the fact that the several families differ widely in their average amount of dark mineral (see page 125 and Fig. 102). Although the average syenite is well within the group of 5 to 50 per cent mafic minerals, the average granite lies close to the lower limit, and the average gabbro and diorite lie close to the upper limit. To divide gabbros or basalts into two groups at the line of 50 per cent mafic minerals would be splitting a fairly well defined group in the region where the group is most thickly clustered; it would put most of our specimens near the dividing line. The prefixes leuco- and mela- are nevertheless valuable, if used by reference to average composition of a family of rocks.

Note 4.—Several recent classifications have proposed a very large number of systematic uniform subdivisions of the main classes, thus attempting to provide a species for every rock, whether known or unknown. The classification of Table III allows subdivisions of the main groups, but goes only a very short way in suggesting rock names for such divisions.

Note 5.—Several minor varieties of rocks have been defined elsewhere as containing certain minerals and certain amounts of these minerals more restricted than the limits given in the table. This is the approved method of progress toward a finally satisfactory classification. Very little agreement has been reached, however, as to the basis of these divisions or how to name them. The term *granite*, for example, is here defined in terms of dominantly silicic feldspar. Some textbooks go much farther and limit the range of *granites* to rocks with dominant orthoclase. It must be recognized, however, that the term has been much used for rocks with abundant albite or oligoclase, and even for many rocks with prominent andesine,—quartz monzonites or granodiorites. The restriction here advocated is suggested as a step of progress but is certainly not a radical change.

Note 6.—The rocks named in the tabular classification are *defined* by their position in the table, and some are not defined in exactly the same way that their names are defined by other published classifications. It is difficult to state all the factors that influenced the choice; some definitions are an average of the prevailing usages, and priority in definition has been respected if not too much outweighed by the prevailing usage.

TEXTURAL, MINERALOGIC, AND SPECIAL VARIETIES IN THE SEVERAL CLANS

The formulas for the several clans are abbreviated but quantitative, corresponding, in general, to the tabular classification (page 50).

x refers to varietal minerals or characterizing accessories.

y refers to microscopic accessories.

z refers to alteration products.

The primary minerals of a rock are thus divided into *essential* minerals, which are required by the species, *varietal* minerals (or characterizing accessories), which distinguish one variety from another, and *microscopic accessory* minerals, which are in so small quantity and so rare as to be ignored in most work on hand specimens. If a mineral is listed in the "formula" of any clan it must be present in notable amount (though occult in the glasses), say about 10 per cent. *When a mineral is said to be "dominant" in a rock or in a portion of the rock, it constitutes more than 67 per cent.* In some rocks varietal minerals or characterizing accessories may be more abundant than the essential minerals specified in the formula of the clan. The microscopic accessories are rarely more than 3 per cent.

Although the scheme is very strictly petrographic, as distinct from geologic, in its attitude toward rock names, certain geologic distinctions have been so much used that the names are included in the tables, so that the student shall not be confused if he finds them in the literature, used in a geologic sense. Chief among these is the use of the term *porphyry* with some compositional adjective, for example, *rhyolite porphyry*, to

signify an intrusive rock as distinct from an effusive, which with the same petrographic character might be called a porphyritic rhyolite.

RHYOLITE-GRANITE CLAN

Formula.—Quartz plus feldspar, dominantly silicic, make more than 50 per cent of the rock, $\pm x \pm y \pm z$.

x = biotite, amphibole, pyroxene, plagioclase, tourmaline, garnet, epidote, magnetite, muscovite (glass)

y = magnetite, apatite, zircon, titanite, pyrite?, fluorite, allanite, cassiterite, ilmenite, fayalite, xenotime, monazite, rutile, calcite?

z = sericite, kaolinite, chlorite, epidote, iron oxides, carbonates, leucoxene, urallite, zeolites, bauxite, quartz, opal

A. Textural and Structural Varieties.

	Rhyolite tuff and breccia (see pages 116 to 119).....	surface beds
	Rhyolite obsidian = hyalorhyolite	
	Other rhyolite glasses (see page 113)	
	Rhyolite perlite	
	Rhyolite pumice	
	Spherulitic rhyolite	
	Aporhyolite = devitrified obsidian	
	Vesicular rhyolite	
	Rhyolite (felsite), aphyric rhyolite	usually surface flows
Aphanites	Porphyritic rhyolite = phyric rhyolite	
	Rhyolite vitrophyr	
	Rhyolite felsophyr	
	Banded rhyolite = eutaxitic rhyolite	
	Rhyolite porphyry = quartz porphyry	
	Granophyr = micropegmatite	usually intrusive
	Granite porphyry	
	Granite	
	Porphyritic granite	
	Orbicular granite	
	Pegmatite	
Phanocrites	Aplite	usually plutonic
	Graphic granite	
	Gneissoid granite (trachitoid granite)	
	(Granite gneiss) (is metamorphic)	

In addition to its strict significance as a rock in this family with a felsitic texture, *rhyolite* is loosely used for all *aphanites* in the clan.

Felsite may be used as a field term to include felsitic light-colored rocks of any composition.

B. Mineralogic Varieties. (Prefix the **x** and even the **y** and **z** minerals.)

Normal granite (biotite)

Leuco- and melagranite

Alkalic granite

Sodic granite

Potash granite (rich in potash feldspar and biotite)

Feldspathic granite**Quartzose granite** (33 to 67 per cent)**Silexite**, (over 67 per cent quartz)**Binary granite** (quartz and one silicic feldspar alone)**Alaskite** (but little dark mineral, includes many textures)**Greisen**¹ (quartz and white mica, altered from granite)**Luxullianite**¹ (quartz and tourmaline, altered from granite)**Protogin** (sheared old-looking granite)

Note that the best terms to indicate mineralogic varieties are those in which the mineral name is prefixed to the rock names of the clan either compounded or in adjectival form.

Note that many of these terms emphasizing one mineral may be combined with textural or structural terms to make a complete descriptive name; thus, sodic rhyolite porphyry, garnetiferous porphyritic granite, biotite rhyolite vitrophyr. Such terms as these are most to be recommended.

*C. Special and Locality Terms.***Aphanites****Liparite** (English synonym of rhyolite)**Nevadite** = granite porphyry

Ailsyte (micropertthite, riebeckite)

Alsbachite = aplitic nevadite with garnet

Comendite (sodic, acmite)

Dahamite (paisanite with albite)

Eurite (felsite; or aphanite of this family)

Grorudite (acmite dike rock)

Krablite (nevadite, tuffaceous?)

Leopardite (manganese stain)

Lithoidite = felsitic rhyolite

Masanite (phenocrysts plagioclase, groundmass orthoclase)

Paisanite (riebeckite)

Quartz bostonite (low quartz. See dike rocks)

Quartz keratophyr (sodic)

Quartz pantellerite (sodic)

Taurite (sodic; spherulitic or granophyric)

Tordrillite (alaskite aphanite)

Tsingtauite = nevadite with orthoclase phenocrysts

Phanerites

Arizonite (approaching silexite)

Beresite (quartz-rich aplite)

Charnockite (hypersthene or enstatite)

Ekerite (arfvedsonite)

Elvan (porphyritic aplite)

Granitite (biotite)

Quartz lindsite (aplite low in quartz)

Rapakivi granite (hornblende porphyry; feldspar, orthoclase in middle, oligoclase outside)

Rockallite (aegirite, albite)

Unakite (epidote, little quartz)

Notes on the Rhyolite-granite Clan.—The name granite refers to the granular texture (Fig. 20). The name rhyolite indicates that flow structures and banding are common in the aphanites of this clan.

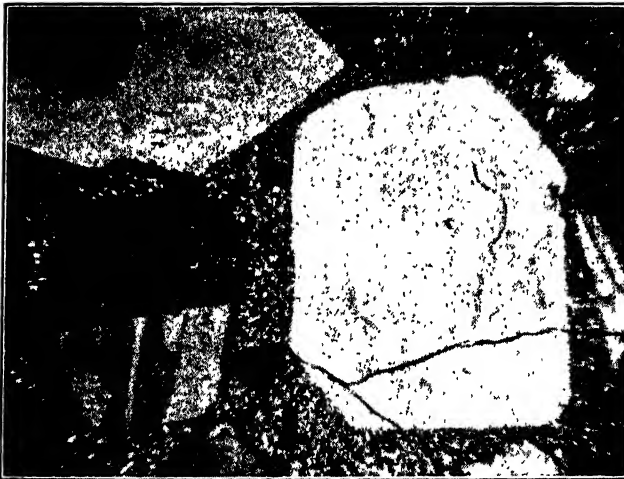
A few of the species and varieties tabulated need further discussion. These will be considered approximately in the order of the table. It is to be emphasized, first of all, that there are no sharp lines between these families, species and varieties. Gradations are characteristic and separation is arbitrary.

¹ Altered so far that formula does not apply.

Rhyolite glasses have indices of refraction less than balsam. Several textural varieties of glass are noted in detail here and not in other families, because the rhyolite glasses are most numerous. The glasses seldom



(a)



(b)

FIG. 22.—Granite porphyry, Saganaga Lake, Minnesota. (a) Hand specimen, about natural size. (b) Photomicrograph, $\times 15$.

occur in large masses and are commonly surface flows and breccias, or selvages on dikes. Real glass is strictly isotropic unless strained. Pitchstones are named from the forms of vegetable pitch—the gum of trees, usually light colored—not from black coal-tar and asphalt pitch.

Vesicular or *cellular* structures appear in lavas charged with gas. Pumice is the result of extreme vesiculation in a silicic glass. When vesicles are filled with secondary minerals the rock is amygdaloidal, but amygdaloidal rhyolites are rare.

Igneous breccias may be of three distinguishable types. First are the fallen fragments formed by explosive volcanic eruptions, usually cemented by ash and vein matter (Fig. 96). Second are flow



FIG. 23.—Porphyritic granite, Rockville, Minnesota. Hand specimen, about one-half natural size.

breccias, formed as a lava flow incorporates and cements fragments resulting from explosions or from breaking of crusts on the flow itself. Third are contact breccias, formed where magma intrudes and cements fragments of wall rock (Fig. 97).

Many *aphanites* show by their chemical composition some constituent not discovered in any recognizable mineral. Such minerals as might be expected but are not seen are called "occult."

Felsitic textures under microscopic examination may show great variety. If the grains are uniform, the texture is *microgranitoid* or microcrystalline, or if extremely fine, microfelsitic or cryptocrystalline. Probably as often the fabric is graphic or spherulitic in felsites and felsitic groundmasses. As a rule the rocks with more than 26 per cent normative quartz have only these textures, and those with less quartz are trachitic. Pilitic and poikilitic textures are rare.

Platy parting is one of the most common structures in outcrops of rhyolite. It may be related to the common banding, eutaxitic structure.

Porphyritic rhyolite is applied to surface flows and *rhyolite porphyry* to intrusives, even if indistinguishable in hand specimen. If the occurrence is not known most men say "rhyolite porphyry." Granite porphyry has more abundant phenocrysts than rhyolite porphyry, but they lie in an aphanitic groundmass. Usually a rock with an abundance of phenocrysts has a coarser groundmass than a rock with relatively few phenocrysts (Fig. 18), but both of these porphyries are aphanites. A phanerite may also show a hiatal fabric and properly be called porphyritic. A *porphyritic granite* has a groundmass that is fine grained as compared to the phenocrysts, but actually too coarse to be aphanitic.

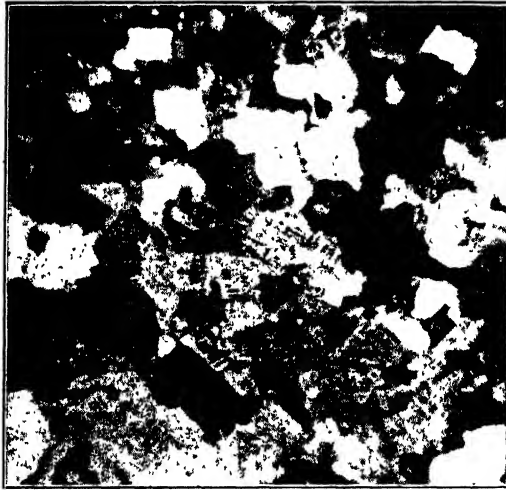


FIG. 24.—Photomicrograph of aplite from Colorado with characteristic sugary texture. Microcline with gridiron twinning, quartz, and orthoclase make up all but a negligible part of the rock. Crossed nicols. $\times 40$.

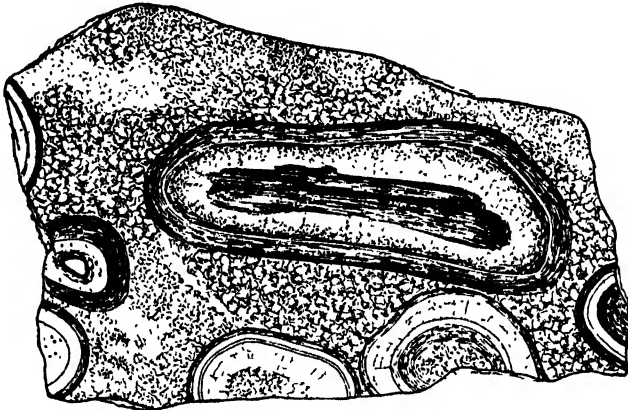
Phenocrysts in this clan are commonly orthoclase, albite and quartz; some of the feldspars are zoned (Fig. 63). In many porphyries weathering loosens the phenocrysts. Some porphyries show fractured phenocrysts, with fragments shaped much like those in tuffs but with lava filling the fracture cracks; such fractures may be due partly to cooling strains and partly to the flow of an immensely viscous magma.

Granophyr is here used as by Rosenbusch for granite porphyries or granitoid rocks in which the feldspars show a micrographic intergrowth of quartz.

The old usage, which distinguished "porphyry" with orthoclase from "porphyrite" with plagioclase, is nearly obsolete; porphyry is a textural term.

Pegmatitic granites have most of their grains as coarse as one centimeter in diameter and from that up (page 69). *Aplites* have most of their grains below one millimeter in diameter and are uniformly sugary (Fig. 24).

Orbicular (Fig. 25) refers to a primary structure; "spheroidal" may be used for the same structure but is better reserved for a secondary effect of weathering (Fig. 203). See the list of further readings. The "orbs" always contain a central nucleus, which is earlier than the outer shells.



(a)



(b)

FIG. 25.—Sketches of polished surfaces of orbicular granites. (After Sederholm.) (a) At Kangasniemi, with wall rock fragment as a core. (b) At Virvik, Borgå, one orb has been worn across the bands and the later bands are "unconformable." About one-half natural size.

Granites having a primary flow structure are perhaps best described as *trachitoid* or gneissoid granites. In these the feldspars and other platy or needle-like minerals seem to have been oriented by movement during crystallization. This leaves the term gneiss and adjective gneissic to indicate metamorphism (frontispiece). Most gneisses are so far deformed as to have augen or mortar structure. Incipient fractures may produce a "rift"—a direction of rock cleavage (see page 39 and Fig. 26).

It has been suggested especially in Europe that magmas may complete their crystallization during deformation, forming real gneisses by a process of "piezocrystallization." The usually accepted criteria of dynamic metamorphism may be found in such gneisses (augen, mortar structure, etc.), but the early development of the structure is indicated by the occurrence of related aplite dikes showing no deformation. These early gneisses are not usually distinguished from gneisses of distinctly later metamorphism. Any large intrusive may occupy a long period, giving time for several alternations of injection and mashing, so that the early phases are deformed by later injections of the same geologic unit.

Alkalic granites, as noted in the general discussion of classification (pages 48 and 49), are probably best defined as those that have minerals

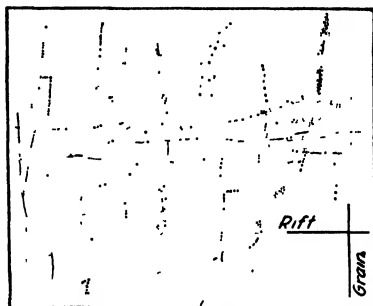


FIG. 26.—Inclusions and minute fractures in quartz in biotite granite, Conway, New Hampshire. (After Dale.) These seem quite sufficient to explain the easy cleavage (rift) of granites in certain directions. $\times 70$.



FIG. 27.—Inclusions of biotite schist in Vermilion granite, Minnesota. In many exposures the schist is so abundant that the granite forms a stock-works rather than a massive intrusion. The granite grades through more and more abundant inclusions, with better orientation, into the schist.

in which the ratio of alkali to silica is greater than 1:6, such as acmite, riebeckite, or some similar mineral. Feldspathoids are naturally improbable in quartz rocks like these (page 31). Most alkalic granites are

notably low in calcic feldspar and have little quartz but some biotite and an abundance of the alkalic feldspars.

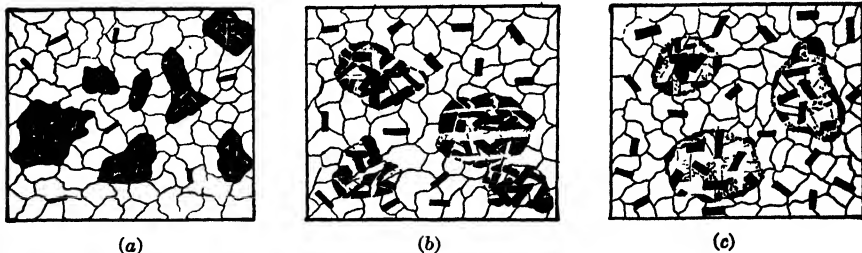


FIG. 28.—Inclusions of greenstone in the Giants Range granite, Minnesota. The sketches show three stages of attack. (a) Fragments recrystallized to hornblende, but not much changed in composition. (b) Fragments largely impregnated with granite. (c) Only ghosts of the fragments remain, and the matrix is notably contaminated with the hornblende of the fragments.

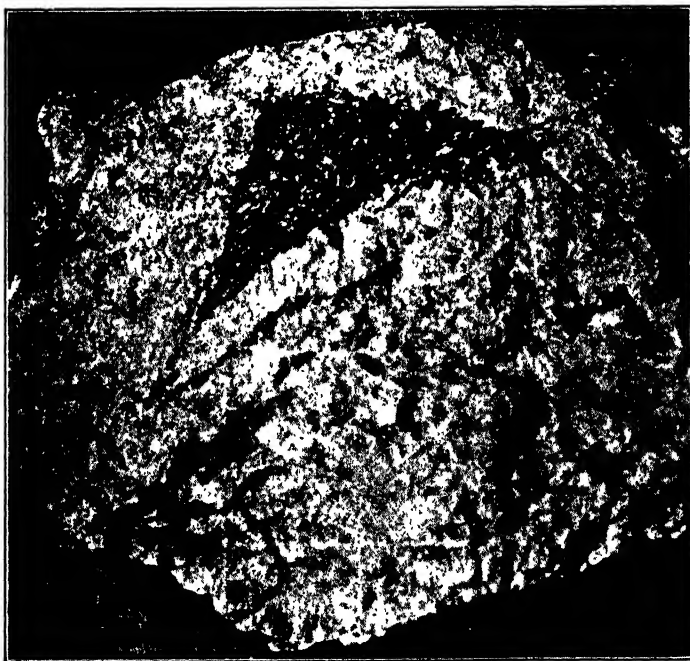


FIG. 29.—Inclusion of biotite schist in granite in the Vermilion district, Minnesota. Slabs of schist seem to have spalled off from the large lump, and crystals have been detached from both. The biotite grains at a distance from the fragment are identical with those near it, and though apparently magmatic they may have been added to the magma from the fragment. About one-half natural size.

The term binary granite is rejected by British petrographers, because it was formerly used to indicate a granite with two micas. A granite with only two minerals is rare, for most of them have two feldspars.

Inclusions in granite are of several kinds; most numerous are fragments of wall rocks accidentally caught in the magma and more or less altered by it (Fig. 27). Some large masses of granite may include in their central parts fragments of more basic rocks, first solidified along their borders—cognate xenoliths.

Granites also commonly contain certain knots, streaks or lumps of darker minerals. It is possible that these also are foreign to the original magma, but as the minerals are identical with those formed early in the rock, it is believed that the spots result from some obscure process of segregation or secretion. There are, furthermore, in many granites hazy



FIG. 30.—Photomicrograph of granite thin section showing biotite, quartz and feldspar. Plain light. $\times 25$.

ill-defined streaks for which we use the German term *schlieren*. These, in the granite, are mostly dark streaks which may result from incomplete mixing of an assimilated fragment (Fig. 262), though some may be a result of segregation and late movement of the magma.

Miarolitic cavities occur more commonly in granites, especially in pegmatites, than elsewhere. They are openings, rarely more than a few inches across, lined with euhedral crystals coarser than the grain of the rock around them but for the most part of similar minerals. These cavities may be partly a result of the shrinkage that occurs during crystallization, but in some the minerals lining the cavity have concentrations of water, boron, fluorine etc., showing that mineralizers collected in the space and possibly helped enlarge the cavity.

Jointing in this family is not exceptional, although rhyolite has a very pronounced tendency to platy parting.

Minerals of the Granite Clan.—Most granite has from 5 to 15 per cent of dark silicates (Figs. 20*a* and 30), an average of about 25 per cent quartz, and the rest more or less equally divided between potash feldspar and silicic plagioclase. Although this average may be taken as a standard

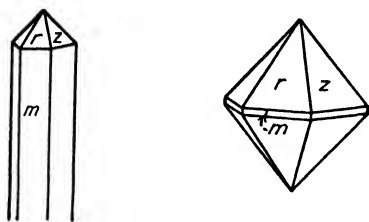


FIG. 31.—The forms of quartz. In porphyries nearly all quartz grains are about equimensional. The elongated crystals of vein quartz such as are common in mineral collections are not common in igneous rocks.

for reference, many granites depart widely from the standard, the rocks with low silica commonly having less orthoclase than plagioclase. The mineralogic variety names serve to distinguish some of the phases, but possibly some method of systematically dividing the group may come into general use later. Leucogranite should mean lighter than average, and melagranite probably over 25 per cent dark minerals. The order of crystallization in most rocks of this clan is the normal or Rosenbusch order (page 44).

Quartz in the formula of this clan is intended to cover the high-temperature forms of silica, cristobalite, and tridymite, which appear in some aphanites, usually surface flows. Cristobalite occurs with feldspar in spherulites but tridymite is said to be more characteristic of late magmatic gas reactions¹ and lithophysae (page 116).

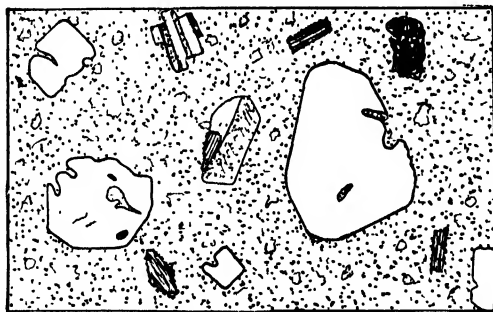


FIG. 32.—Sketch of a thin section of rhyolite porphyry with corroded quartz phenocrysts. Deadwood, South Dakota. $\times 32$.

Quartz is euhedral or corroded in the aphanites and nearly always anhedral in the phanerites. The usual form is pyramidal with a *very short prism* (Fig. 31); if the edges are corroded (Fig. 32) phenocrysts may be deceptively like rounded sand grains or amygdules. Liquid and gas inclusions (Fig. 33) are commonly abundant, as are also microscopic accessory minerals, which make the grains gray or milky or smoky in the

¹ ROGERS, A. F., Natural history of the silica minerals, Am. Mineralogist, vol. 13, p. 80, 1928.

hand specimen. Quartz in euhedral or corroded forms and graphic intergrowths commonly seems uniform in structure, but the anhedral quartz of common granite almost invariably has wavy extinction. This is not limited to metamorphic gneisses but is seen in the freshest of the granites. The strain effects in gneisses are more pronounced and in more nearly parallel bands (see Fig. 226).

Intergrowths of quartz and feldspar have commonly a little over one-fourth quartz, and a mixture of feldspar molecules (Fig. 38). The two minerals appear to have formed simultaneously.

Graphic groundmasses (Fig. 34b) occur more often in intrusives than in flows, but graphic phenocrysts (Fig. 34a) occur in both flows and intrusives. Quartz forms graphic and spherulitic textures only when more than 25 per cent of the rock.

Intergrowths of quartz and silicic feldspar are found in various relations, (1) as graphic structure in phenocrysts, (2) as graphic groundmasses, and (3) as reaction rims in granitoid rocks. Sederholm (see page 243) describes as "myrmekite" a deuteric intergrowth of vermicular quartz

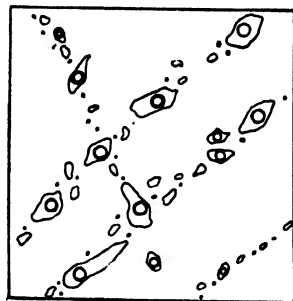
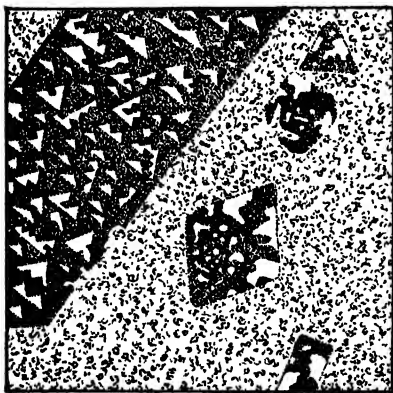


FIG. 33.—Sketch of liquid and gas inclusions in quartz, highly magnified.



(a)



(b)

FIG. 34.—Sketches of graphic intergrowths of quartz and orthoclase in the rhyolite porphyries. (a) Intergrowth in a phenocryst in a partly glassy groundmass. (b) Intergrowth in a groundmass around a feldspar phenocryst. $\times 100$.

and soda feldspar in contrast with graphic intergrowths of quartz and potash feldspar, but in many rocks some vermicular quartz intergrown with untwinned feldspar has replaced plagioclase in a similar deuteric way. The really striking features of the deuteric intergrowth are the rounding of the quartz rods and the apparent encroachment of the inter-

growth on earlier feldspars. Myrmekite appears in rocks wholly unaltered by later metamorphism though some metamorphic intergrowths may resemble it, namely, symplektite with the same minerals and kelyphite with secondary minerals of other sorts (page 361).

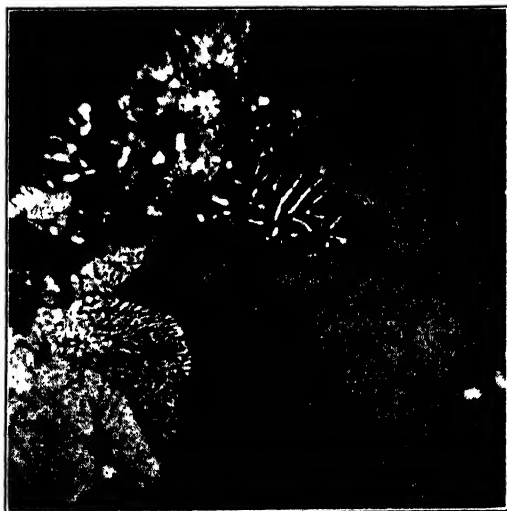
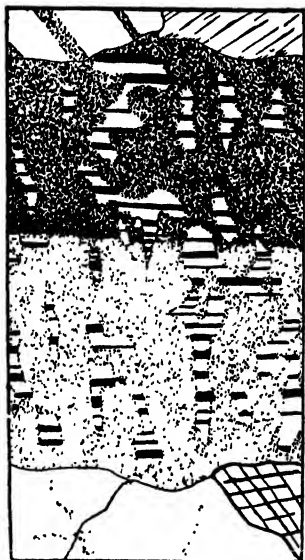
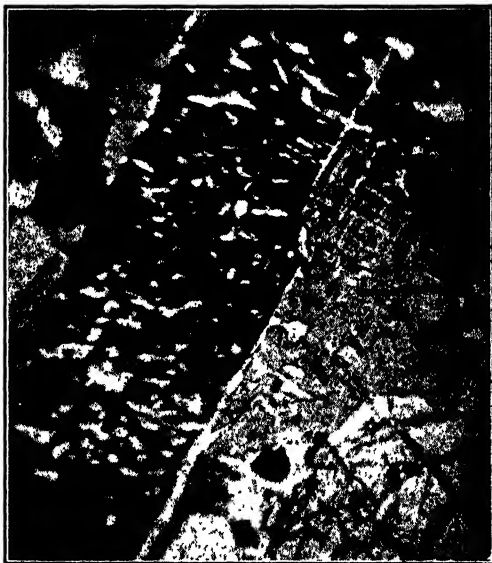


FIG. 35.—Photomicrograph of myrmekite, an intergrowth of quartz and feldspar deuterioally replacing an earlier feldspar. Crossed nicols. X60.



(a)



(b)

FIG. 36.—Perthitic intergrowth of plagioclase in wormy streaks in a carlsbad twin of potash feldspar; orthoclase in the drawing and microcline in the photomicrograph. X100.

The silicic feldspars include orthoclase, microcline, albite and some isomorphous mixtures, as well as the intergrowth, perthite (Fig. 36), which seems to grade into anorthoclase. The glassy variety of potash feldspar, sanidine, is found in relatively recent lavas. Some feldspars may be deuterically intergrown with other feldspars. These several modifications are regarded as equivalents in classification. The feldspars are commonly euhedral and of chunky habit in porphyries and subhedral in granites. Phenocrysts may show some corrosion. Intergrowths with quartz are mentioned above. See also Fig. 38. The twinning characteristic of each feldspar is nearly always present (Figs. 24, 79, 218, and 238). Zonal growth is evident, especially in oligoclase, and especially in the porphyries (see the notes on feldspars of diorites). Orthoclase may have an outer zone of microcline; most microcline is later than other feldspars and fresher, possibly derived from them. Most of the potash feldspars are pink, some others are white, but all are variable. The reddish color in most of these feldspars (but not in all) is probably due to hematite, possibly separated from solid solution.

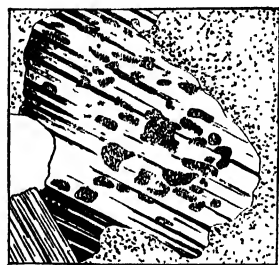


FIG. 37.—Sketch of anti-perthite, small quantities of orthoclase scattered in albite. Crossed nicols. $\times 70$.

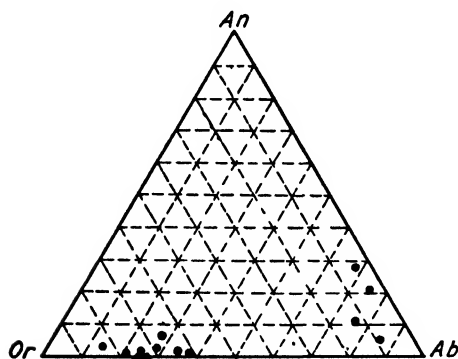


FIG. 38.—Diagram of the proportion of feldspar molecules in the feldspars of graphic granites. (After Bastin.)

They are granites only in a broad sense, as field names. It is best to abandon that loose usage as rapidly as possible.

Biotite is commonly brown and occurs in hexagonal plates or subhedral grains. Extrusive rocks have relatively little biotite, and where it is present there are many signs of reaction with the magma or of resorption. Inclusions of magnetite, apatite, titanite, and zircon are common in the biotite of deep-seated rocks and zircon commonly produces a pleochroic halo in the biotite (Fig. 39). Some biotite may be deuterically intergrown with quartz.

The *amphibole* is nearly always green hornblende, which forms six-sided prisms in the porphyries, and anhedrons in the granitoid rocks. In granites much of the hornblende is altered to biotite, probably by a magmatic alteration or deuteric effect. Riebeckite and arfvedsonite are signs of an alkalic variety of granite.

Pyroxene rarely forms the chief mafic mineral of granites but commonly occurs as crystallites in the aphanites in this family. It may be augite, diopside, or hypersthene; acmite occurs in alkalic varieties. The

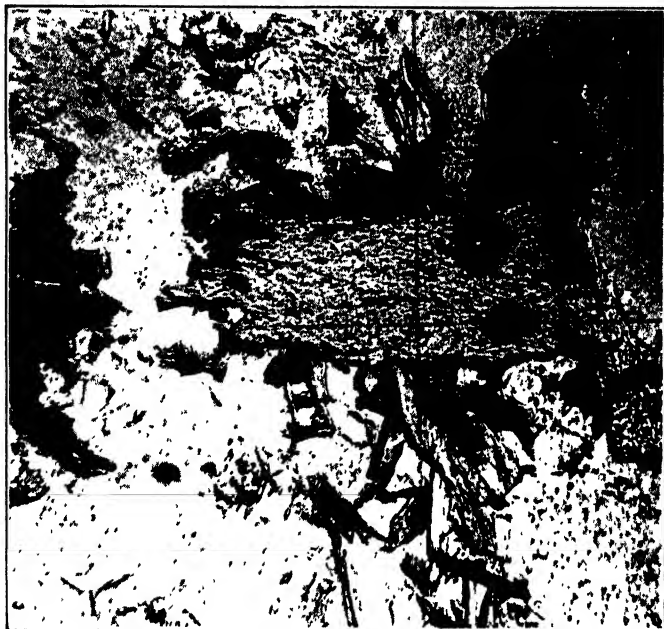


FIG. 39.—Photomicrograph of a thin section of granite in which biotite has been altered to chlorite. In the chlorite, zircon inclusions have caused pleochroic halos, dark shadowy rings, one of them a double ring. $\times 70$.

grains are euhedral in the porphyries but subhedral and commonly altered in granites.

Epidote is commonly secondary but in a few granites it is primary.

Primary muscovite is rare in granites except in the pegmatitic varieties. Even lithium micas occur in these rocks. Garnets, fayalite, topaz, etc., are also found rarely except in pegmatites where the accessory list is very long. Muscovite and hornblende seem to be incompatible in igneous rocks. Biotite may appear with either.

The microscopic accessories, apatite, titanite, zircon and magnetite are very common. Little titanium occurs in ilmenite because in silicic rocks it forms titanite (see Figs. 53 and 83).

It is questionable whether pyrite is truly magmatic.¹ Ore minerals are common in emanations from granite magma into veins and contact metamorphic deposits and seem to accompany solutions of silica, alkalies, and mineralizers, whether early emanations or late mother liquors. In a few granites and granite porphyries quartz and ore minerals seem to be the last products of crystallization of the mother liquor, and to be interstitial to earlier normal granite minerals.

Calcite and dolomite are also questionable primary minerals but seem to be primary in some aplites (see pages 69 to 73).

Deuteric effects in granite result from a reaction of late magmatic fluids with early crystals. In aphanites the chief deuteric effects are the

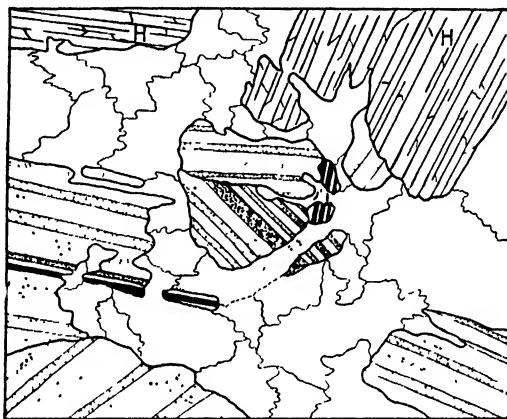


FIG. 40.—Sketch of a thin section of granite, Saganaga Lake, Minnesota. White areas are quartz with sutured contacts; stippled and banded areas are plagioclase, evidently corroded by the quartz; *H*, the hornblende, may also be corroded $\times 25$.

corrosion of quartz and the reaction rims of dusty magnetite and pyroxene around biotite and hornblende; these may be due to the change of conditions at the time of eruption. In granites a greater variety of effects is noted. Early augite and magnetite are commonly bordered or wholly replaced by hornblende or biotite, or by both in succession. Biotite and feldspar may be partly replaced by muscovite. If these replacing minerals are coarse grained they seem deuteric; if fine-grained confused aggregates, they are probably hydrothermal. The late siliceous magmatic residue also attacks the early feldspar, producing myrmekite (Figs. 35 and 40), and a great variety of ore and gangue minerals may at a few places mingle with the replacing quartz. It is not always possible to distinguish sulphides segregated in the magma from those introduced

¹ VOGT, J. H. L., Magmatic differentiation of igneous rocks, Jour. Geology, vol. 29, p. 637, 1921. C. F. TOLMAN and A. F. ROGERS, "Magmatic Sulphide Ores," Stanford Univ. Press, 1916.

later if the later liquids form no characteristic hydrothermal associates (page 110).

Assimilation of clays and other rocks may increase the percentage of mafic minerals (Fig. 41) and may add less common minerals such as sillimanite and garnet. It is questionable if garnet is formed in granite magmas uncontaminated by sediments. Possibly it may be sometimes, as for example in pegmatites.

Alteration in the granite-rhyolite clan may, without shifting the rock into the metamorphic class, change feldspars largely to muscovite,

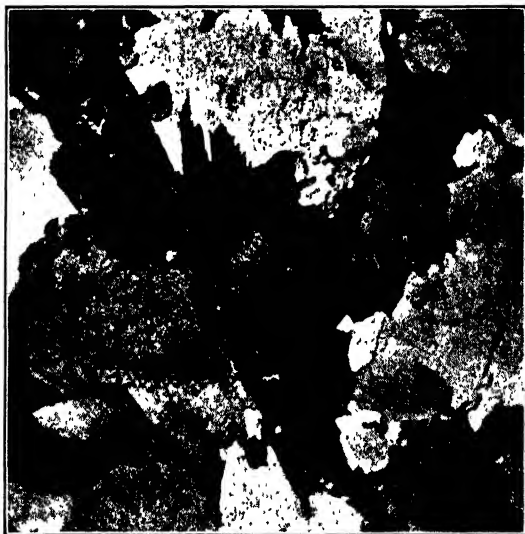


FIG. 41.—Photomicrograph of Vermilion granite with exceptionally abundant biotite (compare with Fig. 30). This occurs near a cluster of inclusions of biotite schist and probably resulted from partial assimilation of the schist. Crossed nicols. $\times 45$.

sericite or kaolin and change ferromagnesian minerals to chlorite, epidote and hydrous iron oxides. Dynamic action may fracture some crystals, bend the biotite, and complicate the twinning of the plagioclase; it seems also to facilitate a change of orthoclase to microcline and of biotite to muscovite.

Contrast of Aphanites and Phanerites.—A contrast may be drawn between aphanites and phanerites in the granite clan. The mafic mineral is mostly pyroxene in aphanites, and biotite or hornblende in phanerites. Biotite in many aphanites is resorbed and has not the zircon inclusions with halos. The feldspar is more commonly zoned in aphanites than in phanerites, but perthite and microcline occur almost exclusively in phanerites. Tridymite and cristobalite occur exclusively in aphanites. The composition of an average aphanite is said to be more siliceous and

more alkalic than the average phanerite of the same clan, but both vary widely. No aphanite grades into siliceous dikes as the pegmatites do.

Pegmatites and Aplites

These rocks deserve a separate description and discussion on account of several unique features. The term pegmatite was given by Haüy in 1822 to the graphic intergrowth of quartz and feldspar, but by 1850 it was in use for such rocks of coarse grain as commonly contained the graphic intergrowth, though that texture was not necessary. Most pegmatites are dikes near the edge of a large mass of plutonic rock to which they are somewhat related in mineral composition (Fig. 42). The group is so variable that exceptions can be found to nearly any definition proposed. "Micropegmatite" does not imply coarse grain but, referring back to the original meaning, implies micrographic texture.

Aplite is a sugary-grained dike rock, commonly related to pegmatite, occurring in the border zones of plutonic rocks and less commonly in the wall rocks. Aplites have much sharper walls than pegmatites, which seem to permeate or be gradational to their walls in many places.

Descriptive Notes on Pegmatites.—1. Pegmatites are characteristically coarse grained but variable. Enormous crystals occur, some as much as 40 feet long, but pegmatites grade into aplite even in a single dike. Some may be banded in gravitatively settled bands (Fig. 45), and a few may be trachitoid. Many of them grade into rocks with granitoid textures, but no porphyry occurs in pegmatites.

2. Graphic intergrowths of quartz and silicic feldspar are common; less abundant are intergrowths of quartz and tourmaline, etc. Perthite is very common.

3. Most pegmatites are dikes (or veindikes); some are segregations, pipelike or irregular; many intrude schists *lit-par-lit* (Fig. 46). In the small injections and *lit-par-lit*, magma has acted like a very thin fluid—

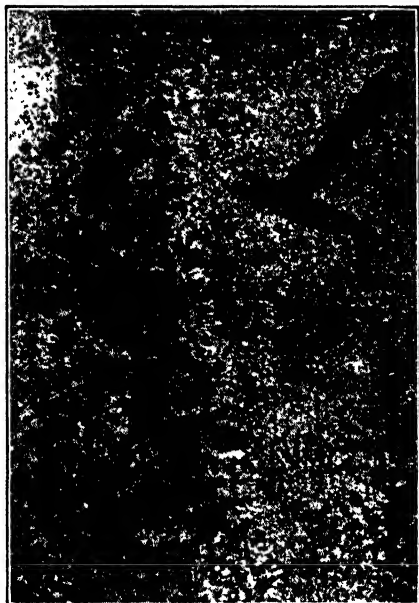


FIG. 42.—Pegmatite dike in granite walls, Minnesota. The dike forms a fairly definite streak but the texture is gradational from the dike to its walls.

there is no crumpling of delicate leaves of schist as by a viscous magma.

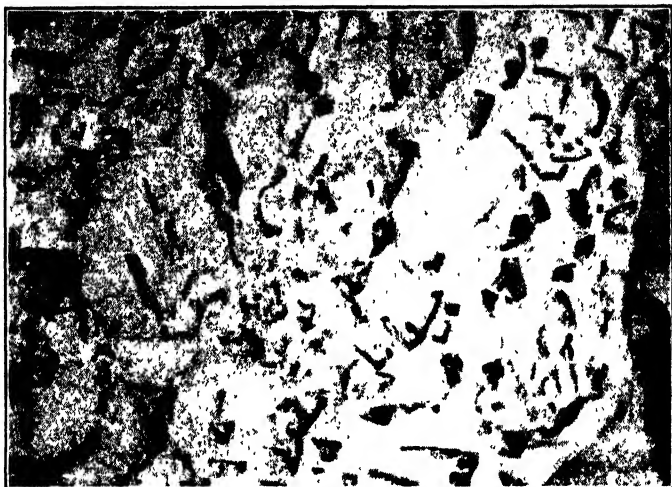


FIG. 43.—Graphic granite from a pegmatite, Lake of the Woods, Minnesota. About natural size.

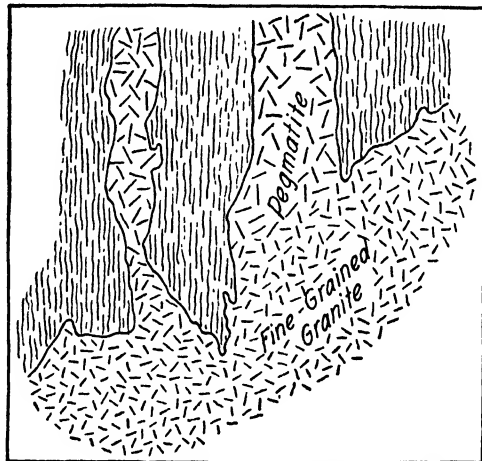


FIG. 44.—Sketch of the contact of a large granite and a schistose wall. (After Bastin.) The granite grades into the pegmatite and more pegmatite than granite injects the wall. Area sketched is about eight feet square.

4. Pegmatites occur near the edges of large igneous masses mostly granite—at least as much outside as inside the masses. It is estimated that about 1 per cent of a large granite magma may evolve into pegmatite and the large masses of pegmatite are almost invariably those that escaped from the main chamber into the wall rock.

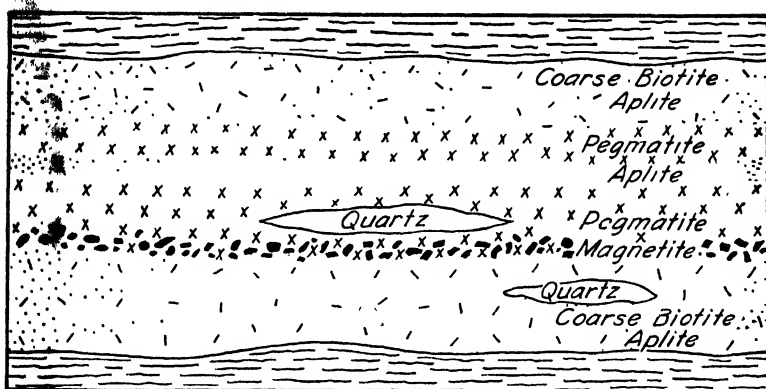


FIG. 45.—Magnetite zone in the pegmatite of the Vermilion batholith. The pegmatite is nearly horizontal and about four feet thick, and the magnetite may have settled a little not growing symmetrically in the dike.



FIG. 46.—A pegmatite dike crossing dark biotite schist and injecting it *lù-par-lù*. About natural size.

5. Granite pegmatites and aplites grade into quartz veins on one hand and syenite pegmatite on the other. Less commonly rocks of other clans have pegmatites, which may be of the same mineral composition as the clan or may grade from that to granite. A granitic composition is assumed for all pegmatites unless otherwise specified. It is not certain whether the quartz veins that grade into pegmatites have anywhere valuable ores of metals like gold or tin. Some aplites related to diabase grade into carbonate veins.

6. The quartz of pegmatites has optical characters indicating that it crystallized at about 575°C.

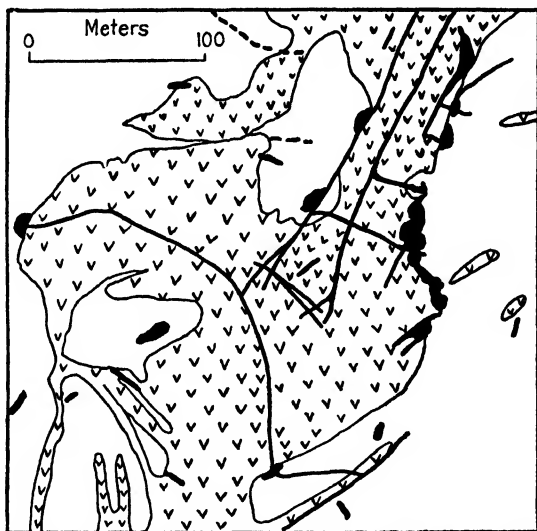


FIG. 47.—Map of a portion of the Matahuala district, Mexico. (After Spurr.) Heavy lines show veins in and near the monzonite. Where the veins reach the adjacent limestone several chimney-like deposits have developed, but few veins extend unaltered into the limestone. Such veins are analogous to pegmatites which connect with bunchy contact rocks in limestone.

7. The order of crystallization is erratic—usually normal in the mass and reversed in miarolitic vugs.

8. There are common signs of replacement, probably deuteric effects.

9. It is estimated that pegmatites are more siliceous and have less dark silicate than the massive rock to which they are related. Few are sampled for analysis, however, because they are so coarse and variable that a fair sample is not easily obtained.

10. There are remarkable concentrations of rare elements, the earths and mineralizers, beryl (Be) in crystals up to a ton, spodumene (Li) logs 40 feet long, tourmaline (B) and topaz (F) masses, apatite (P) veins, etc. Columbite and tantalite are hardly known except from pegmatites.

11. There are signs that water was abundant in some pegmatite magmas; namely, muscovite replacing some feldspars; biotite much more commonly than the less hydrous hornblende and augite; liquid inclusions in the minerals; miarolitic cavities; gradation to veins; absence of pegmatites in rocks formed near the surface, where pressure would not be enough to hold in the gases; veinlike bands and comb structures; thorough differentiation as if thinly fluid, as for example from gabbro to granite or feldspar to quartz; great penetrating power as if not viscous; contact action; some mineralizer was needed to keep the quartz fluid to about 575°C.

The water is estimated as 10 to 75 per cent. In some pegmatite magmas there was possibly not much water, if one may judge from the lack of contact action and miarolitic cavities.

12. The fine grain and sharp walls of aplites indicate that they probably formed from magma that was not highly hydrous, possibly from a pegmatite magma that reached a position where water could readily escape. This suggestion of a genetic relation between the extra coarse and extra fine rocks is supported by a number of occurrences of the two rocks in a single dike. Where pegmatites and aplites are related to granites the pegmatites are said to be richer in quartz and in orthoclase than the aplites.

The difference between aplite and pegmatite is well shown by the behavior of dikes where they pass from igneous rock to limestone. A sharp-walled dike of aplite may pass with little change from one rock to the other or at most become scattered in the more easily broken limestone. Pegmatites, on the contrary, may be well defined dikes in the igneous rock and at the limestone wall may form a chimney or bunch of carbonate-silicate rock such as is usually attributed to contact metamorphism. The more hydrous magma reacts with the limestone at once instead of crossing it for a distance in a fracture (see Fig. 47).

No aphanites are known to be related to pegmatites, though some speculation may be indulged in as to cherts related to lava flows.

TRACHITE-SYENITE CLAN

Disregarding the small content of mafic minerals, the rocks that consist mostly of alkali, alumina, and silica form feldspars, if there is about 65 per cent silica; those with more than 68 per cent will have quartz; and those with less than 63 per cent will have nephelite. Rocks with neither quartz nor nephelite will therefore be restricted to a relatively small range in silica, whereas the granites and nephelite-syenites range more widely. The variable mafic minerals extend the range a little, but it is to be expected that the syenites will be a comparatively small group of rocks.

Formula.—Feldspar, dominantly silicic, no other felsic mineral over 5 per cent, $\pm x \pm y \pm z$

x = olivine, corundum, etc., as in rhyolites and granites.

y = quartz, etc., as in rhyolites and granites.

z = serpentine, etc., as in rhyolites and granites.

A. Textural and Structural Terms.

Aphanites	Trachite tuff and breccia.....	surface beds
	Trachite obsidian and other glasses	
	Trachite ¹ (felsite), aphyric trachite	
	Porphyritic or phyrice trachite	
	Trachite vitrophyr	usually surface flows
	Trachite felsophyr	
	Vesicular trachite	
	Trachitic trachite	
	Trachite porphyry	
	Rhomben porphyry (rhombic phenocrysts of feldspar)	usually intrusive
Phanerites	Syenite porphyry	
	Syenite	
	Porphyritic syenite	
	Aplitic syenite	usually plutonic
	Pegmatitic syenite	
	Gneissoid syenite (Syenite gneiss)	

B. Mineralogic Varieties. (Prefix the x , and even the y and z minerals.)

Normal syenite (hornblende)

Alkalic syenite and trachite

Sodic syenite and trachite

Shonkinite (over 50 per cent mafic)

Leucosyenite (over 95 per cent felsic)

C. Special and locality terms.

Aphanites	Phanerites
Pantellerite , sodic	Minette (mica, see dike rocks)
Keratophyr , sodic	Vogesite (hornblende, see dike rocks)
Bostonite (persalite, anorthoclase, see dike rocks)	Nordmarkite (albite, perthite and a little quartz)
Absarokite (olivine and augite phenocrysts)	Akerite (sodic, approaching monzonite)
Domite (chiefly oligoclase and altered sanidine)	Albitite (albite)
Hyalotrichite (glassy)	Cascadite (olivine, augite, minette)
Kaiweike (phenocrysts, albite with border of orthoclase, some olivine)	Durbachite (much biotite)
Kenyte (alkalic, vitrophyr)	Hedrumite (with accessory nephelite)
Orthophyr (orthoclase porphyry)	Holyokite (albite aplite with calcite)
Ouachitite (biotite in glass dike)	Krageröite (albite rutile aplite)
Sanidinite (contains sanidine)	Laurvikite (perthite and augite)
	Lestiwarite (leucocratic syenite aplite)
	Lindoite (nearly all micropertite)
	Pulaskite (trachitoid alkalic syenites)

¹ The term trachite is also loosely used for all aphanites in this clan.

Aphanites

Selagite (biotite, approaching minette)
 Taimyrite (orthoclase, perthite \pm quartz)
 Verite (biotite, olivine, in pitchstone)
 Volcanite (augite, anorthoclase)

Phanerites

Toensbergite (over 90 per cent alkali feldspar)
 Umptekite (microperthite and arfvedsonite)
 Yogoite (about equal silicic feldspar and mafic)

Note on the Trachite-syenite Clan.—The original “sycnite” was the hornblende granite of Syene on the Nile, but by general custom the term has come to mean rocks without much quartz. Shonkinite is named from an occurrence at Shonkin Sag, Montana. The name trachite refers to the typically rough surface of hand specimens of these rocks. Historically the term *trachite* first meant a light-colored porphyry, later an orthoclase rock; here it is used for rocks with silicic feldspar. *Trachitic texture*, common in the trachites and not rare in other felsites, consists of fine needles or plates in roughly parallel position (Fig. 48). The coarse-grained rocks with similar primary parallelism are *trachitoid* or *gneissoid* (Fig. 49). These terms have only textural significance and should not be confused with compositional terms; but rocks rich in quartz rarely become trachitic in texture.



FIG. 48.—Sketch of trachitic texture in a trachite, Haddington. $\times 100$.

Five porphyritic rocks are distinguished (see page 43). The term porphyritic trachite is used for surface flows and trachite porphyry for intrusives, even if identical in hand specimen. Most men say trachite porphyry, if the occurrence is unknown.

Trachite glasses, being relatively rare, are listed in the table without textural detail (see pages 113 to 116).

Minerals of the Trachite-syenite Clan.—The minerals need little description further than to say that they resemble those in the granite clan. The average syenite has about 20 per cent of dark minerals and 30 to 40 per cent each of orthoclase and albite or oligoclase.

Many feldspar grains are elongated, but others in phenocrysts are nearly equidimensional in common forms (Fig. 50). “Rhombic” feldspars in the rhomben porphyries are commonly anorthoclase or “potash oligoclase.”

Hornblende is abundant and is the dominant mafic mineral in the syenites (Fig. 51). Most hornblende in these rocks is green, but blue sodic amphiboles appear in the alkalic syenites.

Pyroxene in syenite is commonly a light-colored augite (Fig. 52). It grades from greenish diopside in calcic syenite to aegirine in the alkalic



FIG. 49.—Trachitoid (or gneissoid) porphyritic syenite. About one-half natural size.

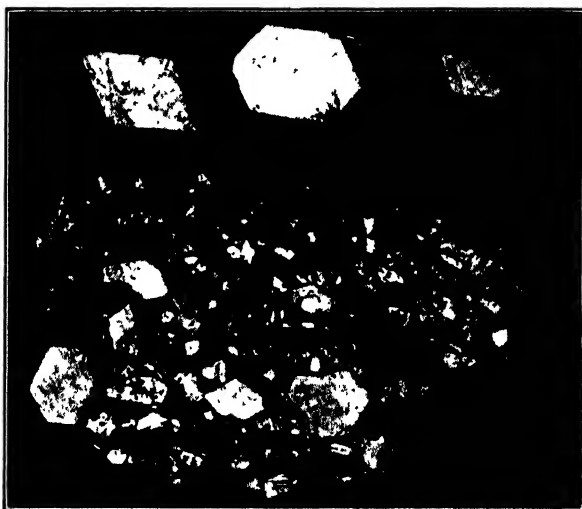


FIG. 50.—Trachite porphyry, Good Springs, Nevada. Slight weathering has loosened the phenocrysts, and a few detached crystals are shown beside the rock. About natural size.

varieties with commonly some zonal growth. Hypersthene is known in some of the aphanites.

Titanite is more common in syenites than in granites and may be visible macroscopically.

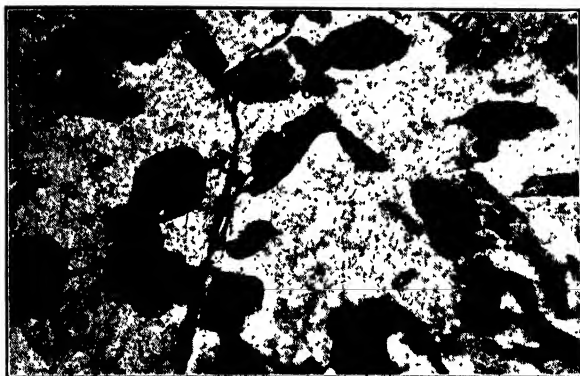


FIG. 51.—Photomicrograph of syenite with euhedral hornblende in coarse feldspar. Vermilion Lake. Plain light. $\times 60$.

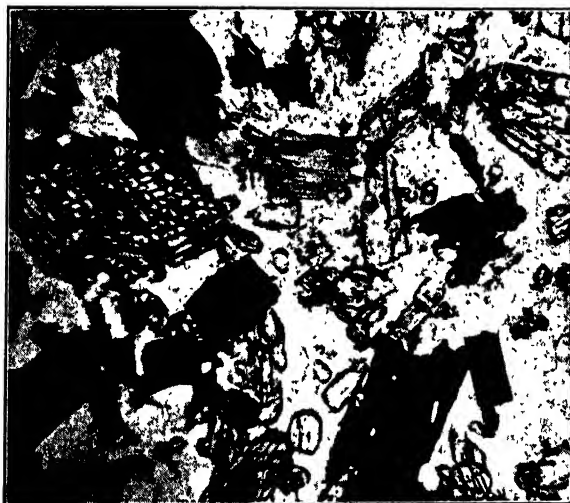


FIG. 52.—Photomicrograph of shonkinite with abundant augite and biotite. Snowbank Lake, Minnesota. Plain light. $\times 25$.

As to structure, texture, order of crystallization, inclusions, jointing, and deuteric effects, what was said of the granite clan is true of the syenite clan and need not be repeated here. There are slight differences between aphanites and phanerites as in the granite clan.

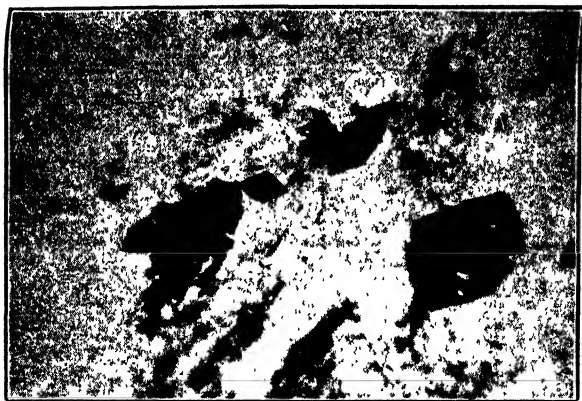


FIG. 53.—Photomicrograph of alkalic syenite with acmite-augite crystals and titanite wedges, in an area of alkalic feldspars. Linden, St. Louis County, Minnesota. $\times 15$.

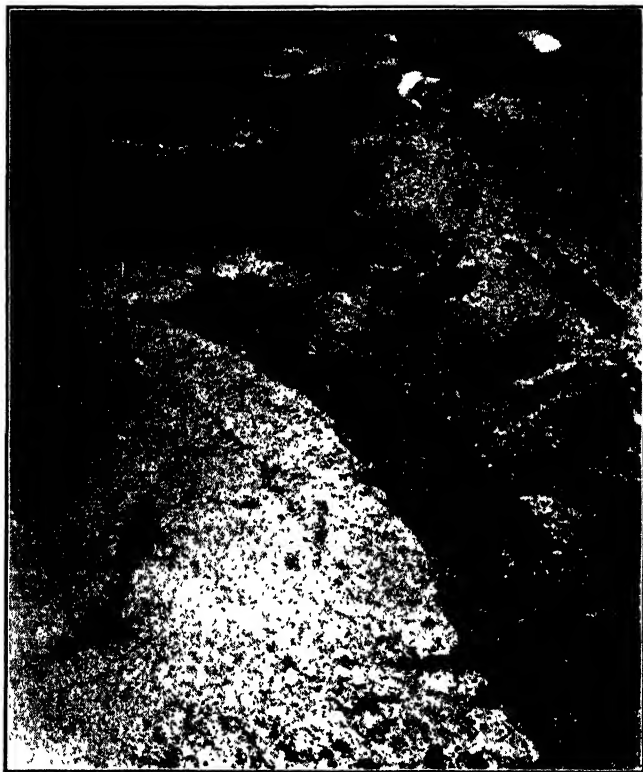


FIG. 54.—Inclusion or schlieren of shonkinite in syenite. Snowbank Lake, Minnesota. The contacts are sharp in some places but gradational in others. The scale is indicated by the folded ruler one foot long.

PHONOLITE-NEPHELITE SYENITE CLAN

Formula.—Feldspar, dominantly silicic + feldspathoid, not more than twice as abundant as feldspar, $\pm x \pm y \pm z$.

x = plagioclase, monoclinic pyroxene, amphiboles, olivine, biotite, etc.

y = titanite, iron oxides, garnet, fluorite, corundum, zircon, carbonates, apatite, etc.

z = carbonates, serpentine, epidote, kaolinite, bauxite, zeolites, chlorite, iron oxides, quartz, leucoxene, sericite, etc.

A. Textural and structural varieties.

Aphanites	Phonolite tuff and breccia	surface beds
	Phonolite obsidian and other glasses	
	Phonolite ¹ (felsite), aphyric phonolite	
	Porphyritic or phyrlic phonolite	
	Phonolite vitrophyr	usually surface flows
	Phonolite felsophyr	
Phanerites	Trachitic phonolite	
	Phonolite porphyry	usually intrusive
	Nephelite syenite porphyry	
	Nephelite syenite	
	Porphyritic nephelite syenite	
	Aplitic nephelite syenite	usually deep seated
	Pegmatitic nephelite syenite	
	Gneissoid nephelite syenite	
	(Nephelite syenite gneiss)	

B. Mineralogic varieties.² (Prefix the **x** and **y** minerals and the particular feldspathoid present.)

Nephelinitoid (phonolite, glass base)

Malignite (syenite over 50 per cent mafic)

Leuco-nephelite syenite (over 95 per cent felsic)

C. Special and locality terms.

Aphanites	Phanerites
Apachite (rich in alkalic mafic minerals)	Agpaite (alkali > aluminum)
Heumite (dike, excess mafic)	Borolanite (garnet and altered nephelite, spotted)
Katzenbuckelite (mica, tinguaitite porphyry)	Canadite (albite)
Kenyte (phonolite vitrophyr)	Chibinite (eudialyte-syenite, sodic amphiboles)
Leucitophyr (leucite phonolite with much leucite and nephelite)	Ditroite (sodalite)

¹ The term phonolite is also loosely used for all aphanites in this clan.

² The British committee suggests that "phonolite" be used to indicate the particular feldspathoid nephelite. Then if leucite occurs with orthoclase and no nephelite they would name it "leucitetrachite"; if with both orthoclase and nephelite, "leucite phonolite."

Aphanites

Madupite (crystals of diopside in glassy leucitic matrix)

Monchiquite (dike, phenocrysts of amphibole, biotite, etc., in glass or analcite)

Pollenite (trachitic phonolite)

Tinguaite (green acmite)

Wyomingite (phlogopite, leucite)

Phanerites

Fergusite (pseudoleucite in malignite)

Foyaite (hornblende nephelite syenite) (or trachitoid acmite nephelite syenite)

Kakortokite (eudialyte-syenite, trachitoid)

Laurdalite (litchfieldite with anorthoclase)

Litchfieldite (gneissoid, albite)

Lujaurite (eudialyte-syenite)

Mariupolite (albite)

Marosite (malignite)

Miaskite (biotitic)

Naujaite (sodalite syenite, poikilitic)

Orendite (leucite, sanidine)

Notes on the Nephelite Syenite Clan.—The term phonolite refers to

the fact that many fragments of these rocks have a sonorous quality when struck.

Malignite was named from an occurrence near the Maligne River, Ontario.

A surprisingly large number of nephelite syenites are trachitoid.

Five porphyritic rocks are distinguished (see page 43).

Minerals of the Nephelite

Syenite Clan.—The average nephelite syenite has about 10 per cent of mafic minerals, but outcrops differ widely. The average rock has also about 25 per cent nephelite, leaving about 60 per cent of

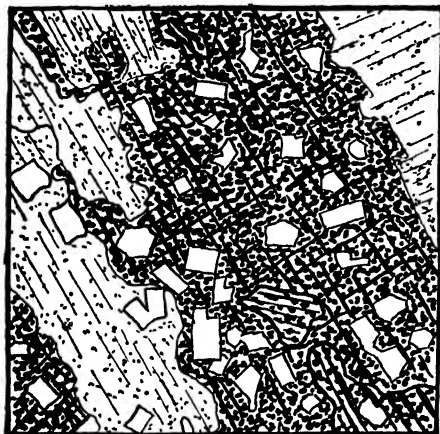


FIG. 55.—Sketch of thin section of phonolite, Devils Tower, Black Hills. Euhedral nephelite in acmite and feldspar. $\times 70$.

potash feldspar and albite-oligoclase; commonly both are present in large proportions and there is much perthitic intergrowth. The feldspathoids may precede or follow the feldspars in time of crystallization.

The feldspars are much like those in the syenite clan, large, platy or needle-like and well oriented.

The alkalic rocks are so peculiar that even 3 per cent of feldspathoid may serve as "essential" in classification rather than the customary 10 per cent.

The feldspathoid is largely gray, brown, or red nephelite with a greasy luster, in both the aphanites and phanerites. The nephelite is euhedral in many aphanites (Figs. 55 and 56), but more commonly anhedral in the syenites, intergrown with orthoclase and replacing albite deuterically. Leucite is more abundant in aphanites than in phanerites. In large grains in syenite it is commonly altered to pseudoleucite or some secondary

mineral (Fig. 57). Leucite is euhedral, and most of it is characterized by zonal inclusions. Commonly if one feldspathoid is present, others may be expected—sodalite, noselite, hauynite (Fig. 58), cancrinite, etc. They alter to zeolites very readily.

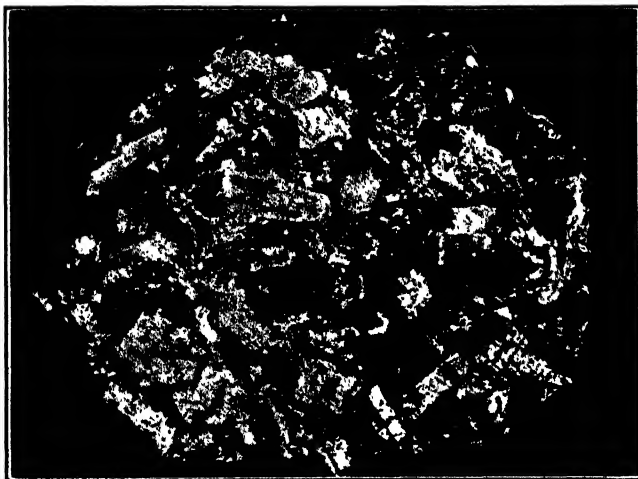


FIG. 56.—Hand specimen of nephelite syenite. Bushveld complex. Nephelite in blocky hexagonal grains, more nearly euhedral than in most phanerites. About natural size.

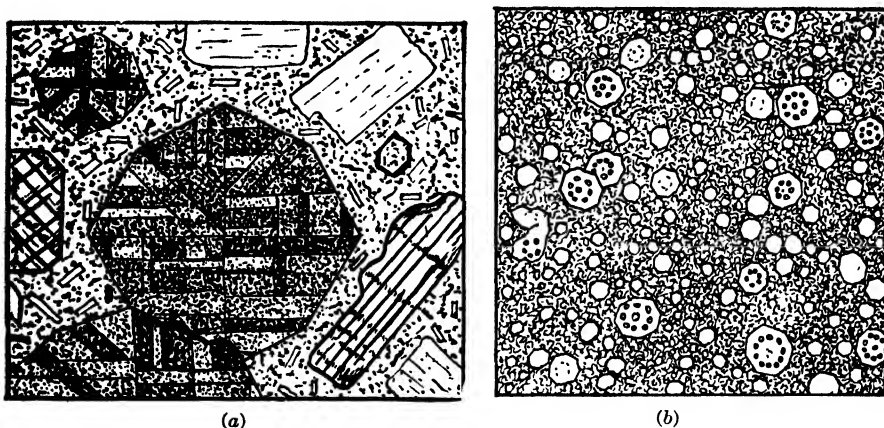


FIG. 57.—Sketches of thin sections of pseudoleucite in large grains, and leucite in small grains in leucite phonolites. The regularly arranged inclusions are characteristic. $\times 60$.

The fact that quartz cannot be present is evident from the formulas (page 31). Rocks deficient in silica are the only igneous rocks likely to have corundum.

The pyroxene is commonly aegirine with zoned and "hourglass" structure. Orthorhombic pyroxenes are rare but are found in some aphanites.

Hornblende is mostly green but in some rocks brown. Sodic amphiboles, blue riebeckite and brown barkevikite are more prominent than hornblende.

Biotite may range to extremes—phlogopite and lepidomelane.

Carbonates are listed with varietal minerals, because in many rocks the grains of carbonate are intergrown with such igneous minerals as olivine and aegirine, so as to appear primary and igneous. Carbonic acid



FIG. 58.—Sketch of thin section of haunite with dusty border zone in a leucite phonolite. X35.

is known to be present in many magmas, and, if it is held under pressure while the magma crystallizes, some of it may not unreasonably be expected to combine with the metals as carbonate as the rock cools. Further, it is supposed by some that the rocks of this clan are originally formed largely by the assimilation of carbonate rocks. These form calcium silicates and settle, leaving a magma enriched in alkalis. With this theory of origin it would be surprising not to find some carbonate relic, possibly so recrystallized as to appear primary. Some experimental

work, however, has cast doubt on the crystallization of calcite from a magma; it melts only at 1340°C. and more than 1,000 atmospheres, unless reduced by fluxes— H_2O , etc., and, some rocks containing supposed igneous carbonate are probably capable of explanation as replacements of igneous rocks by carbonic acid emanations. The criteria for igneous carbonates are still uncertain. If there is some truly igneous carbonate, criteria will be needed to distinguish it from recrystallized limestone relics, and from later replacements, alterations and introductions. See list of further readings.

The black garnet melanite is very noteworthy as an accessory mineral, and may be at least partly a result of contamination.

Deuteric effects are especially prominent in the rocks of this clan. Several of the feldspathoids and associated accessories have fluorine, chlorine, sulphur, carbonic acid and other constituents, suggesting that the formation of the rocks may be in some way related to a concentration of mineralizers. These being late to enter minerals are likely to produce profound alterations in the early minerals as a late magmatic effect. There is commonly good evidence of nephelite replacing albite and of sodalite replacing nephelite.

As noted above there are certain mineral differences between aphanites and phanerites in this clan. Leucite is found only in the aphanites, pseudoleucite in the phanerites and large phenocrysts of the aphanites. Perthite appears only in the phanerites.

LATITE-MONZONITE CLAN

Formula.—Silicic and medium feldspars, each making one-third to two-thirds of total feldspar, no other felsic mineral twice as abundant as feldspar, $\pm x \pm y \pm z$.

x = amphiboles, biotite, quartz, pyroxenes, feldspathoids, etc.

y = titanite, apatite, iron oxides, zircon, allanite, garnet, sulphides, olivine, etc.

z = carbonates, chlorite, epidote, iron oxides, serpentine, quartz, sericite, kaolinite, leucocene, zeolites, etc.

A. Textural and structural terms.

	Latite tuff and breccia	surface beds
	Latite obsidian and other glasses	
	Latite (felsite), ¹ aphyric latite	
	Porphyritic or phyrlic latite	
Aphanites	Latite vitrophyr	usually surface flows
	Latite felsophyr	
	Trachitic latite	
	Vesicular latite	
	Latite porphyry	
	Trachidolerite	usually intrusive
	Monzonite porphyry	
Phanerites	Monzonite	usually plutonic
	Porphyritic monzonite	
	Pegmatitic monzonite	
	Aplitic monzonite	
	Gneissoid monzonite (Monzonite gneiss)	

B. Mineralogic varieties. (Prefix the x and y minerals.)

Normal monzonite? (hornblende?)

Sodic monzonite and latite

Alkalic monzonite and latite

Granodiorite (see diorites)

Kentallenite (over 50 per cent mafic, granitoid)

Theralite (in part) see diorite

Essexite (in part) see diorite

Leucomonzonite (over 95 per cent felsic)

C. Special and locality terms.

Aphanites	Phanerites
Dellenite (quartz latite)	Adamellite (quartz monzonite)
Allochetite (nephelite, porphyritic)	Akerite (sodic monzonite)
Gauteite (andesine, some with rim of orthoclase, dike rock)	Amherstite (andesine antiperthite)
	Mangerite (perthite and andesine)

¹ Latite is also loosely used for any aphanite in this family.

Aphanites

Kullaite (diabase with some microcline)
 Maenaitite (hornblende latite or aplite)
 Shoshonite (andesine in alkalic ground)
 Toscanite (feldspars in siliceous glass)
 Volcanite (augite and anorthoclase, siliceous groundmass)
 Vulsinite (trachitic latite, may have labradorite)

Phanerites

Masanite (phenocrysts plagioclase; groundmass monzonite)
 Routivarite (garnet quartz monzonite)
 Windsorite (aplitic, some quartz)
 Yogoite (about 50 per cent augite, 20 per cent andesine)



(a)



(b)

FIG. 59.—Diagrams traced from photomicrographs, showing the contrast between granitoid texture (a) and monzonitic texture (b). $\times 30$.



FIG. 60.—Photomicrograph of biotite quartz monzonite, Butte, Montana. Biotite altering slightly to chlorite, and feldspar to sericite. Crossed nicols. $\times 50$.

Notes on the Monzonite Clan.—Latite is named from Latium, Italy; and monzonite from Monzoni, in the Alps of Tyrol, where these rocks are well known.

The average monzonite contains about 20 per cent mafic minerals, 40 per cent orthoclase, 40 per cent andesine, and more or less quartz. Monzonites were first defined as having nearly equal contents of ortho-

clase and plagioclase, but now the term is used for an even wider range of rocks than shown here by the formula.

The plagioclase of monzonites crystallized before orthoclase, and the resulting mixture of euhedral and anhedral crystals is the typical "monzonitic" texture, as distinct from the granitoid texture, in which only the mafic minerals are in places euhedral (Fig. 59).

The minerals need no further comment than that they resemble those in the granite and diorite families (Fig. 60).

The plagioclase in vulsinites and banakites is so calcic that they are transitional to the basalt group.

Five porphyritic rocks are distinguished (page 43).

ANDESITE-DIORITE CLAN

Formula.—Feldspar dominantly andesine (or oligoclase), no other felsic mineral twice as abundant as feldspars, $\pm x \pm y \pm z$.

x = quartz, micas, amphiboles, pyroxenes, notably hypersthene, olivine, corundum, feldspathoids, and other feldspars.

y = apatite, zircon, titanite, iron oxides, etc.

z = sericite, kaolinite, chlorite, epidote, carbonates, iron oxides, serpentine, quartz, leucoxene, urallite, paragonite, (saussurite), etc

A. Textural and structural terms.

Aphanites	{	Andesite (and dacite) tuff and breccia	surface beds
		Andesite (and dacite) obsidian and other glasses	
		Andesite (and dacite) ¹ (felsite), aphyric	
		Porphyritic or phyrlic andesite (and dacite)	usually surface flows
		Andesite (and dacite) vitrophyr	
		Andesite (and dacite) felsophyr	
		Andesite (and dacite) glomeroporphyry	
		Vesicular and amygdaloidal andesite (and dacite)	
		Trachitic andesite (and dacite)	usually intrusive
		Diabasic andesite	
Phanerites	{	Andesite (and dacite) porphyry	
		Rhomben porphyry	
		Diorite porphyry	
		Dolerite (medium grained)	
		Diorite	usually plutonic
		Porphyritic diorite	
		Diabasic diorite	
		Orbicular diorite	
		Pegmatitic diorite	
		Aplitic diorite	
		Gneissoid diorite	
		(Diorite gneiss)	

¹ The aphanites of this family are all (loosely) included in the terms andesite and dacite.

*B. Mineralogic varieties. (Prefix the x and even the y minerals.)***Dacite** (quartz andesite)**Normal diorite** (hornblende)**Anorthosite** (over 90 per cent feldspar, granitoid)**Granodiorite** (quartz diorite with 10 to 33 per cent orthoclase)**Theralite** (mafic nephelite diorite)**Essexite** (felsic nephelite diorite)**Propylite** (andesite altered hydrothermally)**Rhomben porphyry** (potash-oligoclase phenocrysts in rhombic forms)**Tephrite and basanite** (see basalt-gabbro clan)*C. Special and locality terms.*

Aphanites

Camptonite (hornblende phenocrysts in andesite, see dike rocks)**Porphyrite** (loose usage, nearly obsolete)**Bandaite** (labradorite dacite)**Boninite** (glassy hypersthene andesite)**Carmeloite** (augite andesite with idding-site)**Cuselite** (augite mica porphyrite)**Esterellite** (quartz microdiorite)**Melaphyr** (altered dark aphanite)**Odinite** (hornblende ground, augite phenocrysts, see dike rocks)**Ortlerite** (hornblende and plagioclase phenocrysts)**Sanukite** (garnetiferous boninite)**Shastaite** (andesine dacite)**Suldenite** (hornblende and plagioclase in very fine andesite)**Toellite** (granophyr andesite with garnet)**Ungaite** (oligoclase dacite)**Volhynite** (hornblende biotite porphyrite)

Phanerites

Tonalite (mica hornblende quartz diorite)**Spessartite** (see dike rocks, hornblende)**Kersantite** (see dike rocks, biotite)**Malchite** (quartz diorite aplite, see dike rocks)**Aleutite** (porphyritic belugite)**Amherstite** (andesine, antiperthite)**Andendiorite** (quartz augite)**Aschaffite** (quartz diorite, dark dike rock)**Banatite** (quartz diorite with orthoclase)**Belugite** (with basic andesine)**Corsite** (orbicular)**Epidiorite** (diabasic gabbro altered to hornblende rock. Metagabbro might be better)**Gladkaiite** (quartz diorite aplite)**Laugenite** (oligoclase diorite)**Lucite** (fine grain, dikes)**Napoleonite** (corsite)**Oligoclasite** (oligoclase diorite) (formerly altered olivine norite)**Ophite** = ophyte (a diabase)**Orbite** (large hornblende phenocrysts)**Ornoite** (hornblende)**Plagiaplite** (diorite aplite)**Plumasite** (oligoclase + corundum)**Yentnite** (scapolite diorite)

Notes on the Diorite Clan.—The andesites were named from the fact that much of the lava in the Andes Mountains is of this composition. The name diorite was proposed to indicate that these rocks could be distinguished from gabbros, which they somewhat resemble. The distinction here used is that based on the difference between medium and basic feldspars; the older distinction by ferromagnesian minerals was advocated before it was known how often augite turns to hornblende.

It would be consistent and systematic to list a separate clan for dacites and tonalites or quartz diorites. They are here placed in this clan, but the term dacite is so familiar that "quartz-andesite" should be avoided.

Most dacites are intrusive porphyries, and relatively few flows have this nature.

Andesites commonly have a texture to be described as hyalopilitic—a felt of crystals in a glassy matrix.

Theralite and essexite, the plagioclase rocks with nephelite, are variously defined by different authors. Essexite was named from Essex County, Mass., but the rocks at the type locality are so variable—some even hybrid—that the definition is loose. Rosenbusch named the

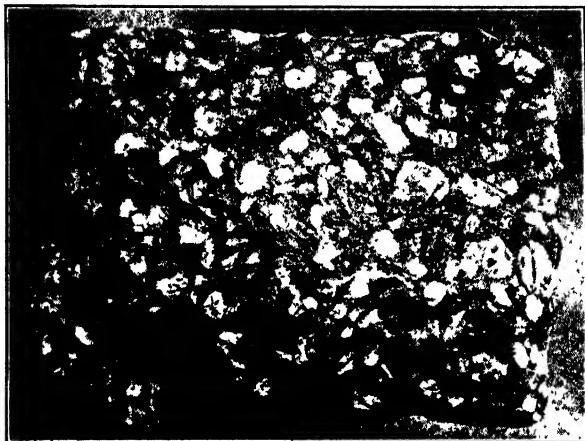


FIG. 61.—Hand specimen of dacite porphyry, Mt. Shasta, California. About two-thirds natural size.

theralites and described them, as having dominant dark minerals like the shonkinites. Hence if essexites are to be distinguished, they may well be restricted to the lighter rocks, those with dominant felsic minerals.

Propylite is an aphanite with plagioclase feldspar, so altered by hot water as to be green and yet be determinable as altered igneous rock. The chief mineral changes are the formation of chlorite, epidote, calcite, pyrite and quartz, largely as pseudomorphs after the original minerals. The term propylite formerly meant andesites of intermediate age, but age is not now used in most classifications.

Granodiorites deserve a special place as a mineralogic variety of quartz diorite because of their abundance. The limits of their variation have been rather strictly defined,¹ but still they constitute a large group.

Dolerites are best defined as rocks of the diorite or gabbro clan with uniform medium to small grains. Many diabases have these medium textures, but diabases have microscopically the ophitic fabric, which is not essential and is commonly absent in a dolerite. Most dolerites are small intrusives, but some may be large flows.

¹ LINDGREN, WALDEMAR, U. S. Geol. Survey, Seventeenth Ann. Rept., Part 2, p. 35 1896; Am. Jour. Sci. 4th ser., vol. 9, p. 279, 1900.

For orbicular diorites see notes on the granites (page 58).

Five kinds of porphyry are distinguished (see page 43), and glomeroporphyries are a result of clustered phenocrysts (Fig. 62).



FIG. 62.—Hand specimen of a glomeroporphyry, Cook County, Minnesota. About one-half natural size.



FIG. 63.—Photomicrograph of diorite from Bannack, Montana, showing zoned crystals. Crossed nicols. $\times 60$.

Amygdaloidal structures are noted more fully under the basalt-gabbro clan.

Minerals of the Andesite-diorite Clan.—The average diorites have about 35 per cent of mafic minerals. The abundant granodiorites have

15 to 20 per cent. As the rocks more than 90 per cent felsic and those more than 90 per cent mafic have special names, the distinction of leuco- and mela-diorites is not very useful for this clan.

Normally the mafic minerals crystallize early, but in certain diabasic varieties plagioclase forms earlier than augite.

Orthoclase is normally in small amounts; if it rises to 10 per cent or more of the total feldspar and has associated quartz, the diorites grade into granodiorites.

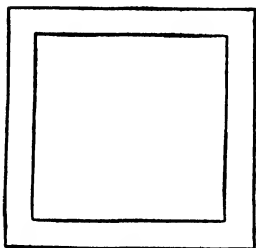


FIG. 64.—Diagram of a zoned crystal. If this is a cross section of a cube the thin outer zone has a volume about equal to the central core.

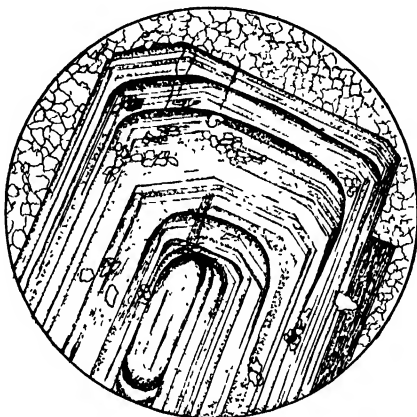


FIG. 65.—Zoned crystal of feldspar from a granite collected by J. F. Wright, Island Lake, Manitoba. The zones alternate many times and there are several "unconformities," probably a result of corrosion. Sketched from a photomicrograph. Crossed nicols. $\times 10$.

Plagioclase constitutes more than 50 per cent of most of the rocks of this clan. It is andesine or calcic oligoclase, usually white, but in places yellow, green or even red. In most rocks of this clan the feldspars are conspicuously zoned (Fig. 63) and have centers more calcic than outer zones; the outer zones, though narrow, may be a surprisingly large percentage of the grains (Fig. 64). The regularity in the zones, moreover, is not universal. An oscillation in zoning is noted in a number of cases and suggests a more or less turbulent condition in the chamber of crystallization (Fig. 65). In the aphanites andesine phenocrysts commonly have inclusions of glass or groundmass. Rhombic crystals of potash-oligoclase in a groundmass give rhomben-porphyry. Dynamic action early causes crossed twinning bands (Fig. 238).

The mafic minerals are usually hornblende, normally green but in some rocks, especially in flows, brown. In the aphanites hornblende and biotite show resorption and corrosion. Hornblende is commonly twinned. The feldspathoid rocks, theralite and essexite, have commonly purplish titan-augite, or green acmite-augite or brown barkvikite.

Augite diorites are less common than hornblende diorites and suggest a gradation toward gabbro, which has usually both augite and a more basic feldspar.

Olivine is not abundant in most diorites. Serpentine formed from olivine has commonly the net of curved markings indicating antigorite as distinguished from that formed from pyroxene and called bastite, if it shows relics of pyroxene structure.

Saussurite is a confused aggregate of hydrothermal alteration products—zoisite, albite, etc.

The other minerals need no comment. Aphanites of this clan are said to be slightly more alkalic and siliceous than phanerites, as is true in most clans.

BASALT-GABBRO CLAN

Formula.—Basic plagioclase, usually the dominant feldspar; no other felsic mineral twice as abundant as feldspar, $\pm x \pm y \pm z$.

x = pyroxene, olivine, micas, amphiboles, feldspathoids, magnetite, ilmenite, quartz, silicic feldspars.

y = apatite, chromite, perovskite, corundum, spinel, garnet, pyrrhotite, iron, graphite, magnetite, ilmenite, etc.

z = chlorite, epidote, carbonate, serpentine, quartz, iron oxides, biotite, zeolites, leucoxene, urallite, paragonite, talc, sericite?, prehnite, albite, saussurite, copper?

A. Textural and structural terms.

	Basalt tuff and breccia.	surface beds
	Basalt obsidian = tachylite	
	Scoria (highly vesicular basic glass)	
	Spherulitic basalt = variolitic basalt	
	Vesicular and amygdaloidal basalt	
	Basalt, ¹ aphyric basalt	
	Porphyritic or phyric basalt	
	Basalt vitrophyr	
	Basalt felsophyr	usually surface flows
Aphanites	Glomeroporphyritic basalt	
	Ellipsoidal basalt = pillow lava	
	Trachitic basalt	
	Diabase basalt	
	Columnar basalt	
	Trap = any dark aphanite	
	Dolerite (medium grained)	
	Basalt porphyry	
	Melaphyr	
	Diabase, and diabase porphyries	usually intrusive
	Gabbro porphyry	
	Lamprophyr, dark dike rock (see page 122)	.

¹ Basalt is also loosely used to include all the aphanites in this clan and other dark-colored aphanites.

Phanerites	Gabbro	usually deep seated
	Porphyritic gabbro	
	Diabase gabbro (ophitic gabbro)	
	Orbicular gabbro	
	Poikilitic gabbro	
	Pegmatitic gabbro	
	Aplitic gabbro	
	Gneissoid gabbro (Gabbro gneiss)	

B. Mineralogic varieties. (Prefix the **x** and even the **y** minerals, except that the feldspathoid minerals must not be prefixed to *basalt*. That combination is used for "no-feldspar" rocks.)

Basalts	Normal basalt and diabase (has olivine)
	Olivine-free basalt and diabase (has not olivine)
	Tephrite (plagioclase + pyroxene + nephelite (or leucite) — olivine)
	Basanite (plagioclase + pyroxene + nephelite (or leucite) + olivine)
	Propylite (see andesites)
Gabbros	Greenstone
	Normal gabbro (augite + basic plagioclase, etc., — olivine)
	Granophyr gabbro and diabase
	Troctolite (olivine and feldspar)
	Norite (orthorhombic pyroxene and feldspar)
	Anorthosite (90 + per cent feldspar)
	Cumberlandite (much magnetite)
	Essexite (see Diorite)
	Theralite (see Liorite)
	Rouvillite (with feldspathoid, felsic)
	Teschenite (with feldspathoid, mafic)

NOTE.—The fact that normal basalt has olivine and normal gabbro has not makes this family complex and hard to arrange in a logical table. One of the worst features of this prevailing usage is that 'olivine-free basalts' have to be named from *what is not present*, a very conspicuous absurdity. It would perhaps be more logical to make normal basalt mineralogically equivalent to normal gabbro, in which case the olivine rocks would be olivine basalt and olivine gabbro respectively. This is not a new suggestion, neither is it established usage, and it would not be understood generally, if the scheme were to be adopted without careful explanation.

The difficulty is increased, of course, by the fact that tephrite and basanite are in this clan and some "basalts" are in a later clan, from which feldspars are absent. It is such confusion as this that tempts petrographers to wholly abandon the older systems that have grown hazy and start anew.

C. *Special locality terms.*

Aphanites	Phanerites
Camptonite (phenocrysts hornblende only) (see dike rocks)	Malchite (quartz gabbro aplitic)
† pessartite (hornblende plagioclase) (see dike rocks)	Euphotide (altered gabbro)
	Aleutite (porphyritic belugite)
	Algovite (any of gabbro family)
	Alivalite (anorthite troctolite)

Aphanites	Phanerites
Odinite (augite crystals, hornblende ground) (see dike rocks)	Belugite (with andesine labradorite)
Aa = aphrolith (highly vesicular, blocky)	Bojite (hornblende)
Pahoehoe = dermolith (ropy surface crusts of lava)	Eucrite (very basic feldspar)
Absarokite (orthoclase)	Forellenstein (troctolite)
Anamesite (dolerite)	Harrisite (much olivine, a little basic plagioclase)
Banakite (labradorite phenocrysts, alkalie ground)	Hyperite (norite)
Bandaite (quartz basalt)	Kedabekite (garnet eucrite)
Ciminite (some orthoclase)	Kyschymite (anorthite, biotite and corundum)
Kulaite (hornblende tephrite)	Ophite (plagioclase formed early)
Melaphyr (altered dark aphanite)	Orbite (hornblende phenocrysts)
Mijkaité (bytownite and augite phenocrysts in red brown basalt)	Ornöite (hornblende)
Navite (olivine dolerite)	Plagioclasite (anorthosite)
Nonesite (orthorhombic pyroxene, and labradorite as phenocrysts)	Soggedalite (diabase, much pyroxene)
Ouachitite (large biotite crystals)	Tilaite (basic plagioclase, much augite)
Pacificite (anemousite)	
Palagonite (brown glass tuff)	
Porphyrite (loose, nearly obsolete)	
Proterobase (old, hornblende)	
Sommaite (orthoclase + olivine \pm leucite)	
Spilite (albitized)	
Tholeiite (vitrophyr)	
Toscanite (orthoclase)	
Vicoite (some orthoclase and leucite)	
Vintlite (brown hornblende and labradorite phenocrysts)	
Weiselbergite (vitrophyr)	

Notes on the Basalt-gabbro Clan.—The term diabase originally indicated a “crossing over” or transition, possibly referring to a cleavage in the rock; but it is now used for rocks in which plagioclase is euhedral with coarse augite filling the interspaces.

Different petrographers use different features to distinguish andesites from basalts. Rosenbusch, Kemp, Pirsson, and Daly describe basalt as having olivine, andesite lacking it. Washington says it is best to restrict basalts to rocks with more than 50 per cent mafic minerals, andesites to those with less than 50 per cent mafic. It is believed, however, that most American petrographers class aphanites bearing andesine as andesite and those with labradorite or bytownite as basalt.

Glasses, tuffs, and breccias (Fig. 96) are summarized on pages 113 to 119). The glasses in this clan are more numerous and notable than in intermediate clans and are given more space in the textural table, not because basalts have much tendency to form glasses, but because flows and small dikes of basalt are very abundant in comparison with aphanites of other clans. Most basaltic glasses are brown or green in thin section and powder but black in hand specimen on account of magnetite crystallites. Their indices of refraction are greater than that of balsam.

The textures of basalts result in a considerable variety of appearances on freshly broken surfaces (Figs. 66, 67, and 68).

Ellipsoidal basalts or pillow lavas have ellipsoids a few inches to a few feet across marked by slight variations in color and texture and alteration (Fig. 69). It is said that the top of an ellipsoid is more likely to be



FIG. 66.—Hand specimens of basalt with fine grain and conchoidal fracture. About one-half natural size.

gently domed and the bottom to fill irregularities in the ellipsoids below it. It is believed that many ellipsoids form under water, especially if erupted in large quantities over wide areas (see the reading lists).

Cellular cavities and amygdulæ are especially common in the abundant basalts of this clan (Fig. 70). They occur mostly near the tops of flows as a result of the rise of gases separated from the lava, but some pipelike cavities near the base probably result from steam rising from the ground water in the rock below the flow. The bending of these near the top may show the direction of flow (Fig. 71). Two contrasting types of tops are noted in basalt flows, based largely on the

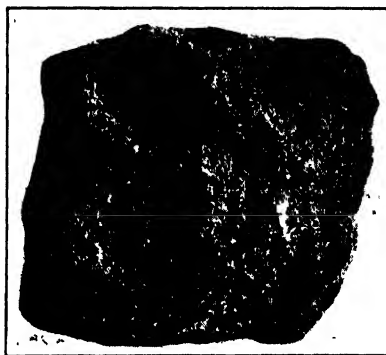


FIG. 67.—Hand specimen of uniform medium-grained basalt. About one-half natural size.

abundance of amygdulæ. In pahoehoe the upper cells may be so abundant as to coalesce forming large cavities under a ropy crust. In aa, which is more cellular, the top may be rough, blocky and fragmental. There are gradations between the two types. These features have some economic significance if ores have been deposited in the cavities, for the fragmental tops have more continuous openings and more variable thicknesses of porous rocks.

Greenstone and propylite are names for altered aphanites of this clan. Greenstone, however, may be extended to cover not only slight hydro-

thermal effects, but dynamic action enough to make the rock really metamorphic—there are greenstone schists.

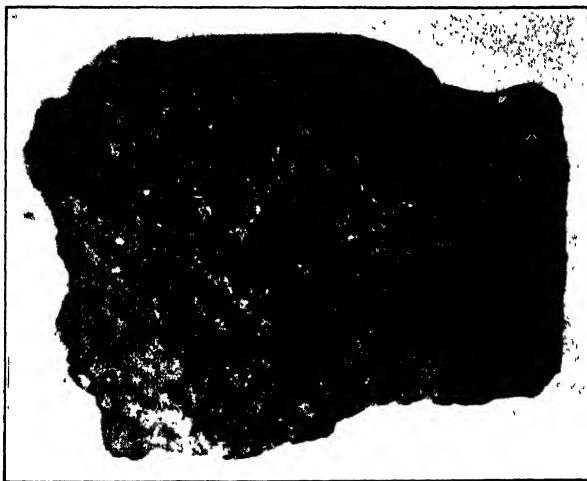


FIG. 68.—Hand specimen of diabase basalt, weathered to show the mottled character of an ophitic rock. About natural size.

Certain aggregates of minerals occur in basalts. These are rounded to angular lumps, often referred to as nodules, mostly of olivine. They are of uncertain origin, probably segregations. Fragments of anorthosite in diabase (Fig. 97) may be of similar origin, related to glomeroporphyry (Fig. 62).



FIG. 69.—Ellipsoidal basalt. Sioux Lookout, Ontario. (Photograph by F. J. Pettijohn.)

Regarding dolerite see the notes on the diorite clan.

In distinguishing the several porphyries see page 43.

Columnar joints are commonly attributed to shrinkage during the cooling that followed solidification (Figs. 16 and 72), but see also the list of further readings on structures and textures.

The needle-like forms of plagioclases in gabbro make this clan especially likely to show a flow structure. Gneissoid gabbros, or trachitoid gabbros or primary gabbro places.

gneisses are found in many

Minerals of the Basalt-gabbro Clan.—Approximately half of the average gabbro is composed of mafic minerals, commonly augite, olivine and magnetite (Fig. 20). The other half is labradorite or bytownite. The group of gabbros, however, shows a notable abundance of rocks that

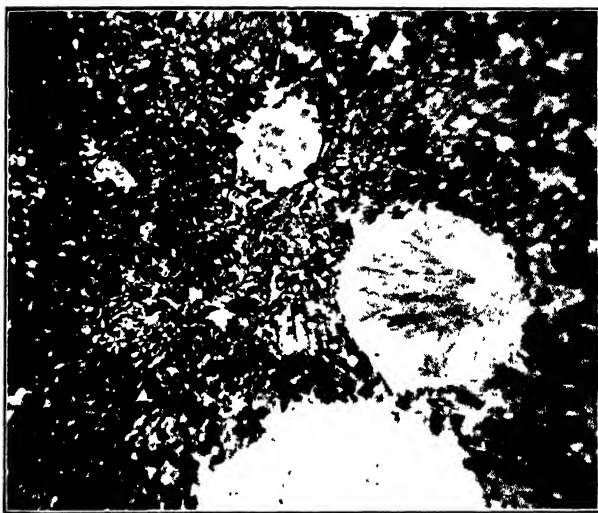


FIG. 70.—Photomicrograph of round amygdules of radial zeolite, and smaller less regular pseudoamygdules of zeolite in a diabase flow. Lake Superior. $\times 25$.

depart widely from the average. Gradations toward anorthosite, magnetite, pyroxenite, and peridotite are common in most large masses of gabbro; and the extremes and gradations are closely intermingled in the field with the near-average kinds of gabbro. Basalts have normally

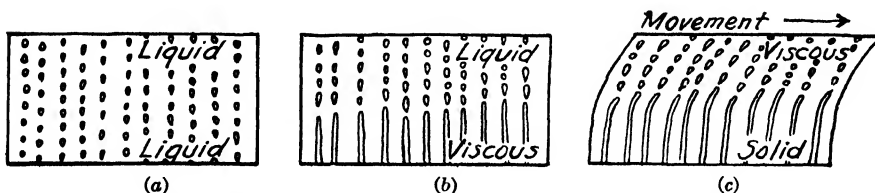


FIG. 71.—Drawings illustrating the development of bent pipe amygdules. (After W. H. Emmons.) (a) Gas bubbles rising through liquid lava. (b) The base of the lava cools to a viscous mass through which the rising steam forms pipes. (c) The base solidifies with pipes in it and the flow of the viscous lava above bends the tops of the pipes in the direction of movement, which is shown by the arrow. Secondary minerals filling the pipes form pipe amygdules.

about 50 per cent of mafic minerals, but show a more limited variation than gabbros.

As the rocks more than 90 per cent felsic and those more than 90 per cent mafic have special names, the prefixes leuco- and mela- are not very useful for this clan.

The normal Rosenbusch order of crystallization is no more characteristic of the gabbro clan than the reverse order. Many of these rocks give evidence that plagioclase crystallized before some of the mafic minerals, notably augite. The resulting fabric, called ophitic or diabasic,



FIG. 72.—Columnar jointing in diabase sill, Grand Marais, Minnesota.

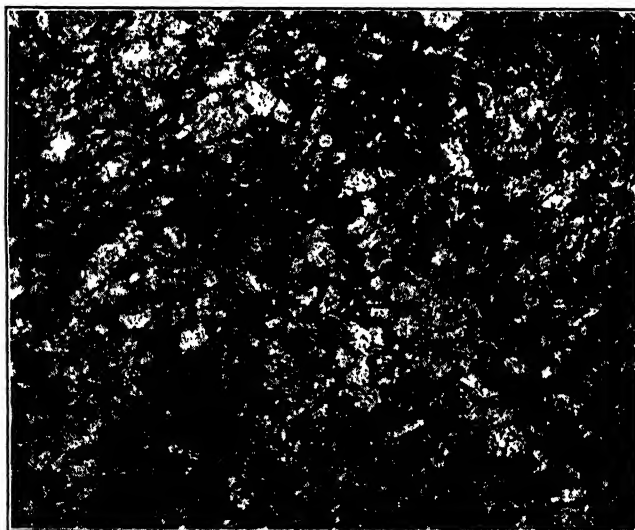


FIG. 73.—Hand specimen of diabase gabbro, Cook County, Minnesota. About natural size.

characterizes the rocks called diabase (Figs. 73 and 74). Vogt¹ notes that ophitic or diabasic textures appear *only when plagioclase makes 55 per cent or more of the rock*, that is, more than the eutectic proportion.

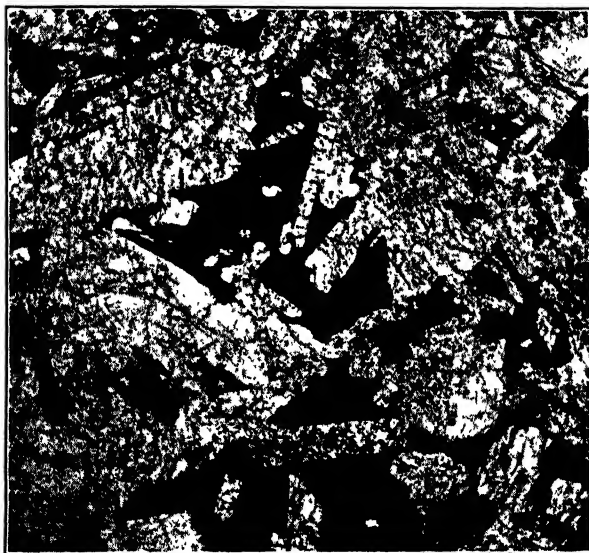


FIG. 74.—Thin section of diabase, Duluth, Minnesota. Both augite (gray) and magnetite (black) formed later than plagioclase. Plain light. $\times 40$.

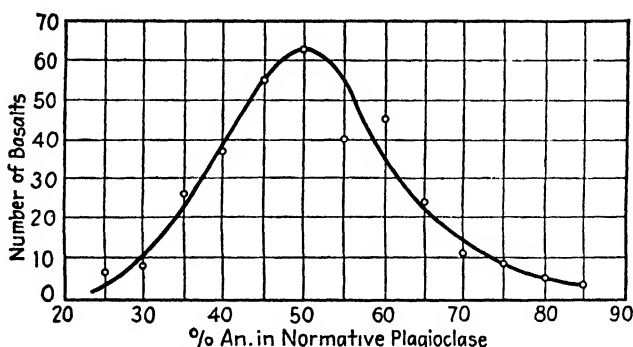


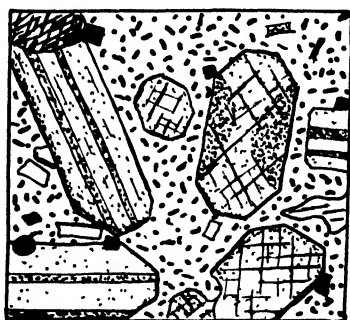
FIG. 75.—Normative plagioclase in 335 basalts. (After Bowen.) The curve has a sharp maximum at 50 per cent anorthite.

As it is primarily a textural variety and refers to fabric rather than to size of grain, the term is probably best used as an adjectival qualifier to rock names, as in diabase basalt and diabase gabbro. Ophite is locally used for diabase flows (Fig. 68) that are mottled in color or luster

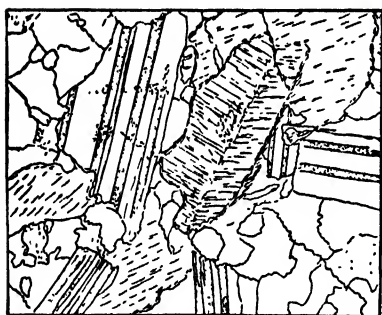
¹ Vogt, J. H. L., Magmatic differentiation of igneous rocks, Jour. Geology, vol. 29, p. 439, 1921.

because of this fabric. Magnetite forms commonly at two stages, part before and part after the plagioclase.

The dominantly basic feldspars have wide twinning bands but are rarely zoned. Crossed twinning is a common early result of stress (Fig. 238). In perhaps half the gabbros the feldspars show euhedral thick tablets characteristic of diabase. In the aphanites the phenocrysts and many of the finer feldspars are euhedral plates. Anorthosite may



(a)



(b)



(c)

FIG. 76.—Sketches of the features of augite in thin sections of basalts and gabbros. (a) Hourglass structure, and slight zoning. (b) Herringbone structure. (c) Schiller. $\times 10$.

have either labradorite or bytownite like other gabbros. Anemousite is a feldspar of the plagioclase series containing in solid solution some carnegite, the triclinic equivalent of nephelite. Basalts with this anemousite feldspar are related to tephrite and basanite.¹

Rarely some silicic feldspar may be associated with labradorite; if so it crystallizes later and is commonly intergrown with quartz, forming "granophyr" (Fig. 78). Albite may also extensively replace the basic feldspar of a basalt, forming "spilite" rocks.

Leucite is found only in the surface rocks, and nephelite is rare in this clan. Analcite has been detected as a groundmass resembling glass, but

¹ BARTH, T. F. W., *Pacificite*: Washington Acad. Sci. Jour., vol. 20, pp. 66-67, 1930.

it has a lower index of refraction than basaltic glass. Melilite is commonly in rectangular grains with peglike inclusions.

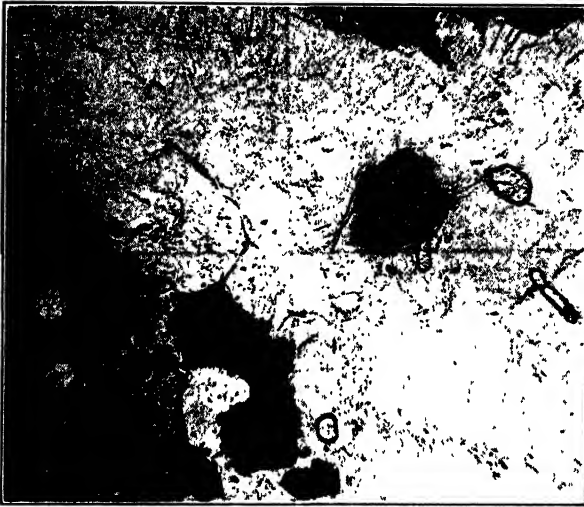


FIG. 77.—Photomicrograph of thin section of garnetiferous gabbro, Granite Falls, Minnesota. Garnet in high relief without cleavage; a small grain of augite cleaved; magnetite, black; and apatite inclusions in feldspar, white. Plain light. $\times 10$.



FIG. 78.—Photomicrograph of thin section of granophyr diabase, Lake Superior. Crossed nicols. $\times 30$.

The pyroxene is augite in a large majority of rocks in this clan; but if the dominant pyroxene in a gabbro is orthorhombic, the rock is called norite. Zoned crystals may be diopside at the early stage, with a border (and related groundmass crystals) of pigeonite, richer in iron and with a

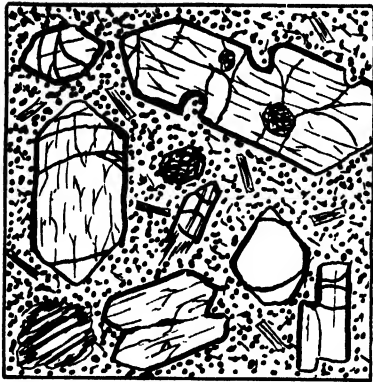
lower optic angle. A purplish color in augite indicates high titanium content. Rocks with diallage, the pyroxene that has an extra cleavage or platy parting diagonal to the prismatic cleavage, were formerly named as distinct varieties, but the differences are slight. The form of the pyroxene is eight-sided stout prisms if euhedral, as in the aphanites. "Hourglass" structure is occasionally noted in augite, as is also a twinning called "herringbone" structure (see Fig. 76). Some alkalic gabbros have acmite-augite. Alteration of the pyroxenes produces corrosion rims of hornblende, biotite, etc., as deuteric effects, and uralite, chlorite, etc., as later hydrothermal effects.



FIG. 79.—Norite, Sudbury, Ontario. (After Dresser, *Econ. Geology*, vol. 12, 1917.) The plagioclase, with albite twinning bands, was fractured before the magma was completely solidified, and the fractures are filled with deuteritic hornblende, sulphides and micropegmatite. Crossed nicols. $\times 70$.

Plagioclase and augite commonly have many dark needle-like or platy inclusions. These have been considerably studied and may be of different minerals in different rocks; perhaps most of them are ilmenite. A lustrous reflection from the included grains is called a "schiller" (Fig. 76c).

The amphibole that forms as a primary crystal in such basic rocks as these is chiefly brown hornblende (basaltic hornblende); it is believed that in the basalts release of pressure on hot hornblende at the time of extrusion causes its dehydration and oxidation and turns it brown. Green hornblende may be deuteritic in gabbros (Fig. 79).



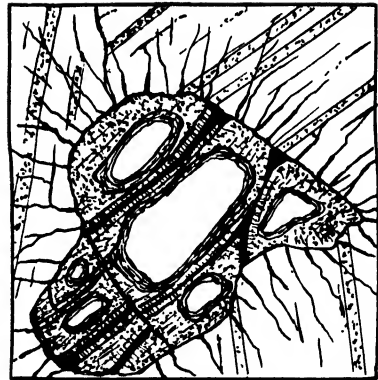
(a)



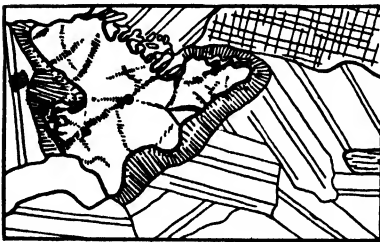
(b)



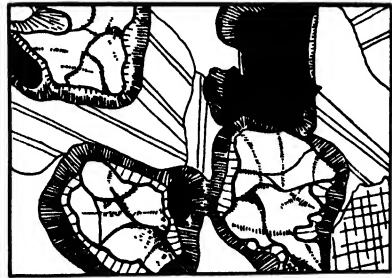
(c)



(d)



(e)



(f)

FIG. 80.—Sketches of the features of olivine in thin sections of rocks of the basalt-gabbro clan. (a) The shapes of phenocrysts. (b) The rounded grains of olivine in poikilitic inclusions in the augite of a diabase gabbro. (c) The network or mesh of serpentine pseudomorphs after olivine. (d) The alteration to serpentine cracks the feldspar around the olivine. (e) and (f) Reaction rims.

Olivine is common and in the porphyries may form euhedral phenocrysts, but in the gabbros it is commonly rounded (Figs. 80a and 80b).



FIG. 81.—Photomicrograph of average olivine gabbro, Duluth, Minnesota. Plain light. $\times 20$.

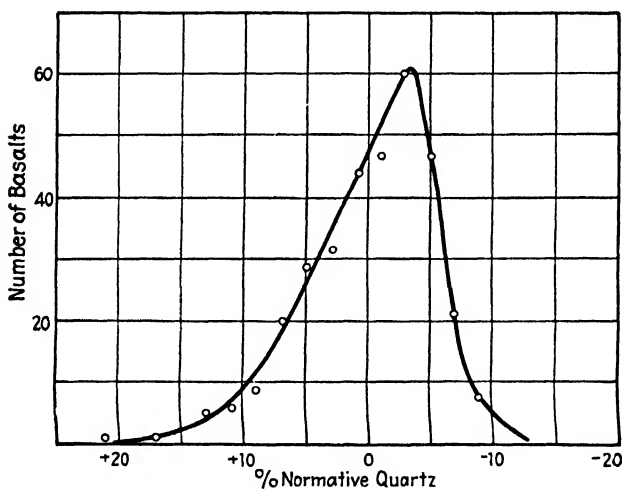


FIG. 82.—Plot of the "positive and negative" normative quartz in 335 basalts. (After Bowen.) Most rocks named basalt by the authors have less than 7 per cent of either quartz or olivine.

Zoned olivines are said to be more magnesian in the core and more feruginous outside. The alteration to serpentine along curved fractures is very characteristic; talc, magnesite, and magnetite may also develop from olivine. The increase in volume in these alterations of olivine commonly produces a series of radial cracks in the adjoining minerals.

When olivine is present quartz is not to be expected, because of the reactions given on page 31. The association is not entirely impossible,

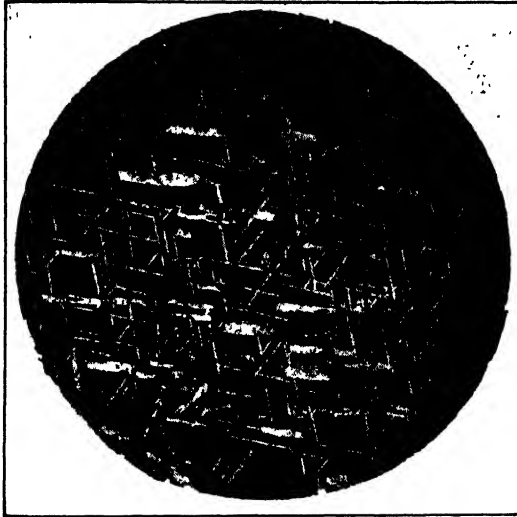


FIG. 83.—Photomicrograph of titaniferous magnetite from a segregation in the Duluth gabbro, polished and etched. (Photograph by G. M. Schwartz.) $\times 100$.



FIG. 84.—Ore minerals primary in the Duluth gabbro. The central prominent gray grain is primary biotite surrounded by plagioclase. The biotite encloses magnetite (black), bornite (smooth gray) and chalcopyrite (rougher gray). Plain light. $\times 15$.

however (pages 244 and 245), and a fair number of gabbros contain interstitial granophyr.

Biotite is not rare but much of it is probably deuterite.

Ore minerals are largely titaniferous magnetite in widely differing quantities and composition (Fig. 83). Cumberlandite is an ore-bearing

rock with perhaps 50 per cent silicates, but it only rarely has feldspar enough to place it in this clan (see pages 106 and 108). Apatite is a common accessory (Fig. 77), and sulphide ores are primary in a number of large masses (Fig. 84).

Reaction rims or coronas are very characteristic of gabbros, especially those of large deep-seated masses. These are somewhat radial, graphic or wormy intergrowths along the contacts of early and late minerals produced by late magmatic or deuteric reactions. The intergrowths are commonly pyroxene and plagioclase, plagioclase and iron oxide, plagioclase and olivine, even biotite and iron oxides. Hornblende and garnet may appear in some. The intergrowths of magnetite and ilmenite are probably deuteric and result from a separation from solid solution.

The minerals of amygdules are commonly calcite, quartz, zeolites in great variety, chlorite and epidote. Copper and silver in amygdules form commercial ore; they occur in some flows that had an early, probably magmatic concentration of hematite, forming a red top on the dark basalt. The iron of these red tops may be increased as much as 40 per cent over that in the rest of the flow, by a process that is probably fumarolic rather than atmospheric.

Hydrothermal alteration in the gabbros commonly attacks the several silicates at once and yields a more complex mixture of secondary fine-grained products than in other common rocks. If the greenish aggregate can not be resolved into its component minerals the aggregate is called saussurite, but the mineral nature of saussurite is probably variable.

Contrast of Aphanites and Phanerites in the Gabbro Clan.—Leucite occurs in the aphanites—tephrites and basanites—but rarely in the gabbros. Basaltic brown hornblende is also much more common in basalts than in gabbros. On the other hand, the extreme concentrations of one of the essential minerals occur only in the phanerites, and there is little tendency for basalt to grade into rocks of monomineralic character like anorthosite, pyroxenite, dunite or magnetite rocks. If the average basalt is found to differ from the average gabbro chemically, the significance of the difference is not at all clear; by definition normal basalt is olivine bearing, and normal gabbro is olivine free; on the other hand, if it is a rule that the average aphanite is more siliceous and alkalic than the corresponding phanerite, it would seem that the difference is fundamentally due to a difference in the conditions under which the two groups solidify.

NEPHELINITE-IJOLITE-MISSOURITE CLAN

Formula.—Feldspathoid, the dominant felsic mineral, $\pm x \pm y \pm z$.

x , y , and z as in the basalt-gabbro clan, *except no primary quartz*.

A. Textural and structural varieties. From felsitic to granitoid.

B. Mineralogic varieties.

Aphanites	{	Leucitite and leucitophyr (no olivine)	}	Usually surface flows and small intrusions
		Nephelinite and nephelitophyr (no olivine)		
		Analcitite and analcitophyr (no olivine)		
		Etc.		
		Olivine leucitite or leucite basalt ¹ (+olivine)		
		Olivine nephelinite or nephelite basalt ¹ (+olivine)		
		Olivine analcitite or analcite basalt ¹ (+olivine)		
Phanerites	{	Etc.		}
		Ijolite (mostly felsic — olivine)		
		Bekinkinite (mostly felsic ² + olivine)		
		Fergusite (mostly mafic ² — olivine)		
		Missourite (mostly mafic + olivine)		

*C. Special terms.*³

Alnöite (dike rock of melilite fergusite porphyry)

Monchiquite (dike, mafic minerals in analcite base)

Ouachitite (biotite monchiquite with no olivine)

Fourchite (augite monchiquite with no olivine)

Coppaelite (porphyritic melilite fergusite)

Euktolite = venanzite

Jacupirangite (ijolite, fergusite and alkaline pyroxenite, usually rich in iron ore)

Monmouthite (hornblende ijolite)

Tawite (sodalite ijolite)

Urtite = ijolite, perhaps approaching fergusite

Venanzite (leucite melilite bekinkinite porphyry)

Notes on the Nephelinite-ijolite-missourite Clan.—These rare rocks are mostly small differentiates with characteristic association. Aside from the association with nephelite syenites, which differ chiefly in having less nephelite, it is noteworthy that they almost everywhere show grada-

¹ The British Committee quite properly object to the use of the term basalt in this clan. The terms "olivine leucitite," etc., are preferable but up to the present time are less used.

² Authorities disagree as to the proportions of felsic and mafic minerals in certain varieties.

³ Not worthy of so much space; the whole clan does not make up one per cent of one per cent of the igneous rocks.

tions into large volumes of less alkalic rocks. Many of them seem to be in the nature of lamprophyrs (pages 121 to 124; see also the list of further readings on alkalic rocks).

Several schemes have been presented for partial or complete classification of alkalic rocks. There are no aphanites equivalent to such rocks as urtite and tawite.

Minerals of the Nephelinite-ijolite-missourite Clan.—The several feldspathoids are commonly associated. See notes on the basalt-gabbro clan. The pyroxene is commonly aegirine but may be purple titaniferous augite in some rocks. Apatite is abundant. Calcite is also abundant but its origin is not clear; it may be partly primary, partly secondary and partly recrystallized from inclusions (page 82). Melilite may be a result of assimilation of limestone.

PERKNITE CLAN

Formula.—Pyroxene, or amphibole, or both; less than 10 per cent felsic minerals, $\pm x \pm y \pm z$.

x = monoclinic or orthorhombic amphibole, mica, etc.

y = magnetite, ilmenite, spinel, pyrrhotite, apatite, garnet, olivine, feldspars, feldspathoids, etc.

z = serpentine, bastite, talc, actinolite, urtite, hornblende, chlorite, epidote, carbonates, quartz, iron oxides, garnet, etc.

A. Textural and structural varieties.

Aphanites	{	Augitite tuff and breccia.....	Surface beds
		Augitite, aphyric augitite	
		Porphyritic or phyrice augitite	
		Augitite vitrophyr	Usually surface flows
		Augitite felsophyr	
		Amygdaloidal and vesicular augitite	
		Augitite porphyry	Usually intrusive
Phanerites	{	Pyroxenite	
		Porphyritic pyroxenite	
		Pegmatitic? pyroxenite (coarse, but not of hydrous magma)	Usually deep seated
		Gneissoid pyroxenite	
		(Pyroxenite gneiss)	

B. Mineralogic varieties. (Prefix **x** and even **y** minerals.)

Hornblendite and amphibolite (more amphibole than pyroxene)

Bronzitite

Hypersthenite

Cumberlandite (much ore mineral)

C. Special and locality terms.

Aphanites	Phanocrites
Monchiquite (phenocrysts of amphibole, biotite, and fewer of pyroxene and olivine + <i>analcite</i> base)	Websterite (both bronzite and diopside)
Fourchite (monchiquite, no olivine)	Eclogite (smaragdite, omphacite, pink garnet)
Ouachitite (biotite fourchite)	Ariegite (highly aluminous)
	Avezacite (hornblende, ilmenite)
	Cromaltite (melanite)
	Diallagite
	Vullinite (diopside, hornblende)
	Yamaskite (hornblende, augite, etc., medium grained)

Notes on the Perknite Clan.—The clan name perknite refers to the dark color of the rocks.



FIG. 85.—Photomicrograph of hornblende. Basswood Lake, Ontario. The rock seems to be a primary segregation from a magma that later formed a large granite batholith. Plain light. $\times 10$.

Pyroxenites and amphibolites occur rarely except as segregations in relatively small bulk related to other intrusives. Intrusives and extrusives are very rare.

Average perknite is more than 80 per cent femic, more than 90 per cent mafic.

The pyroxenes may be orthorhombic or monoclinic or both. Some may be alkalic grading to acmite, especially in the outer zones. The

alteration of pyroxenes in rocks in which they are so abundant as in these commonly yields serpentine (see description under peridotite).

Hornblendite may be deuteric after pyroxenite but is not necessarily so (see Fig. 85).

PERIDOTITE CLAN

Formula.—Olivine and less than 10 per cent felsic minerals, $\pm x \pm y \pm z$.

x = pyroxene, amphiboles, mica, magnetite, etc.

y = magnetite, ilmenite, chromite, pyrrhotite, perovskite, diamond, picotite, spinel, garnet, apatite, corundum, feldspar, feldspathoid, rutile.

z = serpentine (antigorite and chrysotile and bastite), talc, actinolite, urallite, hornblende(?), chlorite, epidote, carbonates, quartz, iron oxides, zeolites(?).

A. Textural and structural varieties.

Aphanites	Limburgite tuff and breccia.....	Surface beds
	Limburgite (aphyric limburgite)	Usually surface flows
	Porphyritic or phyric limburgite	
	Limburgite vitrophyr	
	Limburgite felsophyr	
	Amygdaloidal and vesicular limburgite	Usually intrusive
Phanerites	Limburgite porphyry	
	Peridotite porphyry	
	Peridotite	Usually deep seated
	Porphyritic peridotite	
	Pegmatitic? peridotite	
	Gneissoid peridotite	
	(Peridotite gneiss)	

B. Mineralogic varieties. (Prefix x and even y minerals.)

Cumberlandite (rich in ore minerals)

C. Special and locality terms.

Aphanites	Phanerites
Monchiquite (dike rock, mafic phenocrysts, analcite groundmass)	Picrite (a little feldspar) (see also aphanites)
Picrite (once used for limburgite vitrophyr) (see phanerites also)	Kimberlite (biotitic, porphyritic)
Farrisite (barkevikite, diopside monchiquite)	Dunite (chiefly olivine, chromite specks)
Garewaite (many diopside phenocrysts)	Cromaltite (melanite)
Giumarrite (hornblende monchiquite)	Cortlandite = Hudsonite (hornblendic)
Rizzonite (dikes, limburgite or monchiquite)	Eulysite (wehrlite rich in garnet)
	Harrisite (anorthite, pyroxene)
	Harzburgite = Saxonite (rhombic pyroxene)
	Josefite (dike rock, serpentinized)
	Koswite (rich in diopside and magnetite)
	Lherzollite (diallage and rhombic pyroxene)
	Schriesheimite (hornblende picrite)
	Scyelite (poikilitic, hornblende, biotite)
	Valbellite (hornblende, hypersthene)
	Wehrlite (diallage)

Notes on the Peridotite Clan.—Peridotite was named from the olivine (peridote) that it contains.

Peridotites occur in a few places as independent intrusives, and in a considerable number as small segregations in gabbro magmas.

Minerals of the Peridotite Clan.—Olivine forms early and is commonly rounded. It may make up only a small per cent of the rock, but in dunites it dominates to the exclusion of other minerals except accessory chromite.

Serpentine is the common alteration product, but talc and iron oxides are also notable. The varieties of serpentinous products may be worth distinction.¹ Chrysotile is characterized by a fibrous structure; antigorite by a "mesh" structure; bastite has a structure suggesting origin from a pyroxene. The serpentine may be deuteric or hydrothermal.

Peridotites are of special interest on account of the occurrence of diamonds, platinum and other segregations or enrichments related to them. Diamonds are said to have *grown* as crystals in the magma. Some dikes in the diamond-bearing kimberlite are largely carbonate, and it is possible that carbonate is assimilated and should be listed as a magmatic mineral.

The other minerals are much like those of gabbro or pyroxenite.

ORES

Formula.—Iron (and other) ore minerals usually dominant, $\pm x \pm y \pm z$.

$x + y$ = magnetite, ilmenite, pyrrhotite, chromite, corundum, rutile, hematite and the minerals characteristic of the gabbros, pyroxenites and peridotites

z = alteration products of basic rocks, in particular serpentine

A. Textural and structural varieties.

Granitoid

Trachitoid (see Fig. 17)

B. Mineralogic varieties. (Prefix the x and even y minerals.)

Magnetite rock is most common

Cumberlandite (about 50 per cent ore minerals)

C. Special and locality terms.

Jacupirangite (magnetite and pyroxene, often nephelite)

Krageröite (rutile and plagioclase)

Nelsonite (ilmenite, apatite, rutile)

Urbainite (rutile, ilmenite and sapphirine)

Yamaskite (ilmenite, pink augite, brown hornblende)

Magmatic ores occur mostly as small segregations in gabbro masses or more rarely in small dikes. In some the iron oxides have crystallized later than the plagioclase. See the tabulated criteria for magmatic segregation. Some sulphide ores that are properly magmatic may have

¹ The complexity of serpentinous products is shown by a study in *Am. Jour. Sci.*, vol. 12, p. 515.

been introduced into the silicates at a late magmatic stage (deutericly). These are with great difficulty distinguished from introduced sulphides related to other and later intrusions at high temperatures. The more common late introductions, however, are associated with hydrothermal effects (see also the notes on ore minerals in the gabbro clan, page 104).

CRITERIA SUGGESTING MAGMATIC SEGREGATIONS

Definitions.—Magmatic differentiation is the process by which a magma or molten rock stuff of more or less uniform composition splits up into bodies of different composition. Segregation is a differentiation resulting in a concentration of some particular mineral or element. Some deuteric effects are included here, but pegmatites are not.

1. Minerals

ORES		GANGUE
Magnetite	Platinum	Minerals of igneous rocks
Ilmenite	Diamond	Quartz
Spinel	Monazite	Feldspars
Cassiterite	Graphite	Pyroxenes
Pyrrhotite	Chromite	Olivine
Chalcopyrite	Hematite	Micas
Pyrite	Bornite	Zircon
Molybdenite	Violarite	Garnet
Lollingite	Pentlandite	Amphiboles
Arsenopyrite	Sperrylite	Apatite
Corundum	Niccolite	Feldspathoids

Absence of secondary silicates produced by pneumatolytic and hydrothermal processes.

2. *Structures.*—Igneous structures only. Many crystals euhedral. Some uncrustified banding.

3. *Textures.*—Textures of granitoid igneous rocks. Reaction rims. Complex intergrowth of gangue and ore minerals. Crystalline textures, some coarse and some even porphyritic. Inclusions of sulphides in silicates and corrosion of silicates by sulphides. No corrosion of the sulphides. Euhedral silicates in an ore groundmass. No vugs but a few microlitic cavities.

4. *Sequence of the Ores.*—Apatite, magnetite, hematite, (pyrite?), pyrrhotite, pentlandite, chalcopyrite, bornite, zircon, titanite.

5. *Mineral Combinations and Relations.*—Mostly in basic and differentiated rocks

ORES	ASSOCIATED ROCKS
Apatite and molybdenite	Alkali-rich syenites
Bornite	Hypersthene rocks, norite, anorthosite, diorite
Cassiterite and tungstates	Granite
Chalcopyrite	Gabbro, norite, diabase
Chromite	Peridotite, usually serpentinized
Corundum	Syenite, nephelite syenite, anorthosite, peridotite
Diamond	Peridotite
Ilmenite	Gabbro
Magnetite	Syenite, alkalic syenite, anorthosite, gabbro
Pentlandite	Norite, diabase, gabbro
Violarite	Norite, diabase, gabbro
Palladium	Peridotite
Platinum	Peridotite
Sperrylite	Norite, diabase, gabbro

6. *Associations.*—Same minerals as in the associated igneous rock, but in different proportion. Pyrite, chalcopyrite and bornite. Never pyrrhotite and bornite together. Pyrite and magnetite. Ores with heavy silicates. Magnetite and ilmenite.

7. Field Relations.—Rudely tabular, ellipsoidal or irregular. Schlieren. In sills, laccoliths, etc., in many near the base or border. Grade into igneous rocks. In any one locality ore has a marked preference for one type of country rock. Many deposits contain angular fragments of wall rock. Never related to solution channels. Little migration into the country rock. Not usually hydrothermally altered.

8. Ore Introduction.—Ore introduction into other minerals came between the formation of high-temperature silicates and later hydrothermal alteration.

9. Negative Criteria.—Pneumatolytic minerals. Structures other than igneous. Crustified banding, pseudomorphs. Evidence of metamorphism or metasomatic replacement developing sericite, carbonates, chlorite, urallite, garnets, epidote, bleaching or kaolinization.



FIG. 86.—Meteoric stone, Winnebago City, Iowa, showing the blackened outer surface with characteristic depressions resembling finger marks. About one-half natural size.

METEORITES

1. *Elements in Order of Abundance.*—Fe = 55 per cent, O = 17 per cent, Si = 9 per cent, Mg = 7 per cent, Ni = 5 per cent, no other so much as 1 per cent.

2. *Classification.*

A. *Aerolites.* (Stone.)

1. Ca, Al minerals } textures fragmental,
2. Mg minerals } granular and chondritic

B. *Siderolites.* (Stony iron.)

Iron with Mg minerals

C. *Siderites.* (Iron) The iron may be shown by etch figures to be partly octahedral, partly cubic, and partly massive.

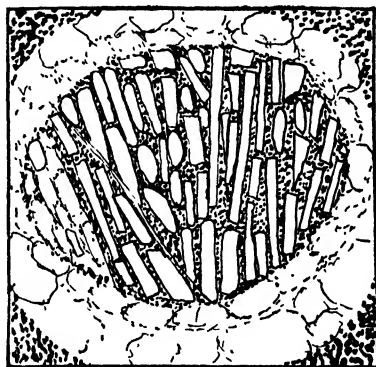


FIG. 87.—Sketch of a thin section of a chondrule in a meteoric stone. $\times 60$.

3. *Peculiarities.*

- A. Black crust and pitted surface (Fig. 86).
- B. Abundance of iron
- C. Chondrules (Fig. 87)
- D. Rare minerals
- E. Etch figures on the polished iron (Fig. 88).

4. *Minerals.*

- A. Terrestrial: Olivine, enstatite, augite, anorthite, labradorite, oligoclase, apatite, magnetite, chromite, pyrite, pyrrhotite, breunnerite, graphite, diamond. Combinations resemble gabbro and peridotite rather than granite.



FIG. 88.—Meteoric iron, polished and etched. Carlton meteorite. About one-half natural size.

- B. Rarely if ever terrestrial: Iron nickel alloys (Kamacite, taenite, plessite), schreibersite (Fe, Ni, P), troilite (FeS), daubreeelite (Fe, Cr, S), lawrencite (FeCl₂), osbornite (Ca, Ti, O, S), cohenite (Fe, Ni, C), moissanite (C, Si), oldhamite (CaS), maskelynite (glassy plagioclase), asmanite (SiO₂)

5. *Textures.*

- A. All igneous, fragmental, tuffaceous in some, almost diabasic in a few. The interstitial glassy material is feldspathic.
 - B. Chondrules (peculiar to meteorites), spherical, many radiate, mostly of *enstatite* and *glass*. Question if they are frozen drops of glass? Abraded later? Most likely a product of tuffaceous origin. Very common.
 - C. Later alteration obscures the tuff and chondrule structures by recrystallization.
 - D. Textural relation of metallic iron indicates *introduction*.
 - E. Black coating is glassy.
 - F. Veins and slickensides appear in some.
- See the list of further readings.

SUMMARIES OF CERTAIN IGNEOUS ROCKS

It may be instructive to summarize certain points in the general outline already given. The rocks covered by a few of the summaries may be tabulated almost as if they constituted a distinct family or clan, but they are really collected under the special heading from many clans. Such tabulations are possible for glasses, tuffs, breccias, lamprophyrs and perhaps others. Summarizing that can not be so readily tabulated covers additional information as to forms, ores and economic materials, the historical geology of rocks and rock series and provinces.

THE GLASSES

Formula.—Dominantly glass \pm the several minerals of felsitic rocks.

A. Textural and structural terms.

Obsidian
Vesicular obsidian
Scoria
Pumice
Perlite
Vitrophyr
Ash
Ash obsidian (glass fragments in a glass)
Obsidian breccia
Pele's hair
Apobsidian (devitrified glass)
Spherulitic and variolitic glass.

B. Compositional terms. (Best to prefix the name of a felsitic rock of same composition to the name of a textural variety.)

Obsidian (acid)
Tachylite (basic)
Pumice (acid)
Scoria (basic)
Spherulitic (acid)
Variolitic (basic)
Pitchstone (acid, H_2O high, "pitchy" luster and color)

NOTE.—Combine the textural and compositional terms, as for example, basalt obsidian, rhyolite vitrophyr, trachite perlite.

C. Determination of composition.

- a. Use index of refraction: table below and curves of Fig. 89.
- b. Minerals: May be used but are not reliable
- c. Color: Light-colored glasses are silicic; some are remarkably clear
Green and brown glasses in section and powder are usually basic
Black and red, owing to inclusions, differ in composition
- d. Specific gravity: As good as index but hard to estimate
- e. Magnetism: Most common in basic glasses
- f. Hardness greater than 7 suggests pitchstone

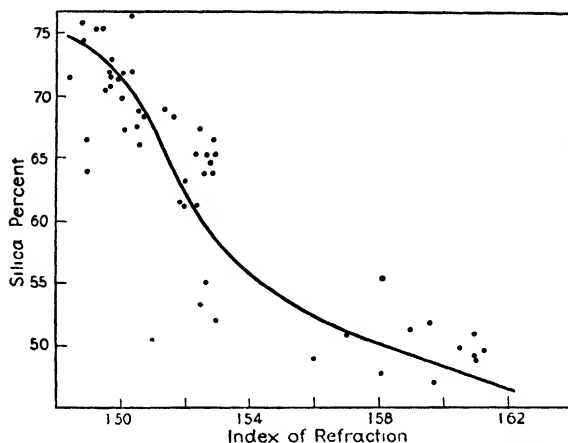


FIG. 89.—Curve of silica per cent and index of refraction of natural glasses. (After George, *Jour. Geology* vol. 32, p. 365, 1921.)

Collector's or author's name of rock		Average index of refraction	Range recorded
Basic Medium Silicic	Obsidian or rhyolite glass	1 492	1 48 -1 51
	Pitchstone	1 500	1 492-1 506
	Perlite	1 497	1 488-1 506
	Pumice	1 497	1 488-1 506
	Dacite	1 511 -	1 504-1 529
	Trachite	1 512 -	1 488-1 527
	Andesite	1 512 -	1 489-1 529
	Leucite tephrite	1 550	1 525-1 580
	Tachylite, scoria, diabase and basalt glass	1 575 -	1 506-1 162
	Palagonite		

D. Locality terms and special names.

Liparite, rhyolitic glass
 Palagonite, basic, brown (altered), in tuff
 Moldavite, meteoric? clear
 Sordawallite, borders of basalt dikes
 Weiselbergite, microporphyritic dolerite vitrophyr.
 Verite, basic, high in phlogopite and biotite
 Tordrillite, alaskite glass
 Mareconite, perlitic rhyolite glass
 Volcanite, high in anorthoclase and augite, chemically similar to dellenite

E. Special features of the glasses.

Mineral dust
 Crystallites (trichites, belonites, globulites, margarites, etc., Fig. 91)
 Gas bubbles or vesicular structures
 Spherulites, variolites, axiolites
 Flowage structure
 Perlitic cracks (Fig. 90)
 Phenocrysts: Commonly feldspars of the sort characteristic of aphanites of such composition or a little more calcic. Quartz occurs in some and olivine in others.
 Devitrification: Gives much quartz and orthoclase, because most obsidian is silicic. Silica may be *added*, however, during mineralization.
 Lithophysae: The minerals of glass cavities—stone bubbles (Fig. 92) are attributed to the effect of mineralizers. Fayalite, quartz, anorthite, biotite, tridymite, garnet, topaz, hornblende, zircon.

Glass is strictly isotropic except under rare strains, but natural glasses commonly contain many crystallites and much dust. As they result from quick chilling of lava, no large masses are known; flows and the thin selvages of dikes are the chief occurrences. The glass may alternate



FIG. 90.—Sketch of perlitic cracks in glass. $\times 15$.

with felsitic bands or form the groundmass of porphyries, the glass being usually more silicic than the associated crystalline material.

It seems that in the course of geologic ages nearly all glasses become crystalline. Few pre-Cambrian glasses have been found unaltered.

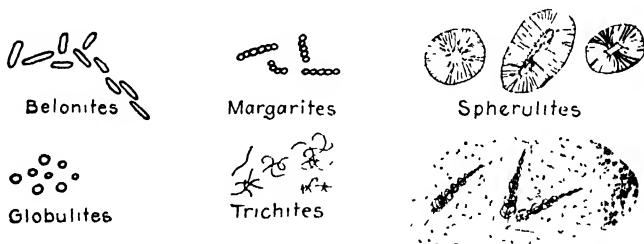


FIG. 91.—Crystallites of several kinds $\times 150$, and spherulites $\times 25$, as seen in natural glasses under the microscope.

The presence of perlitic cracks, or spherulites (Fig. 91), or lithophysae (Fig. 92) is taken as a sign of glassy original textures, even if no glass remains. Spherulites and lithophysae are crystalline, but these aggregates grow only in glasses. The concentric cracks in lithophysae may be partly from cooling and partly from gas set free when spherulites form. Spherulites may form after perlite cracks, for some occur in rows along the curving cracks.

Magmatic water is retained by most glasses, probably because it had not time to escape. Obsidian has about 1.5 per cent, pumice, and perlite about 3 per cent, and pitchstone 5 per cent or more of water.

Analcite may be confused with glass in a groundmass. It has an index 1.47 but is associated with minerals of a *basalt*, in which glasses would be expected to have an index greater than 1.56.

Rarely some flinty crushed rocks grade into an isotropic material that may be glassy (page 413) but varies widely in composition with the

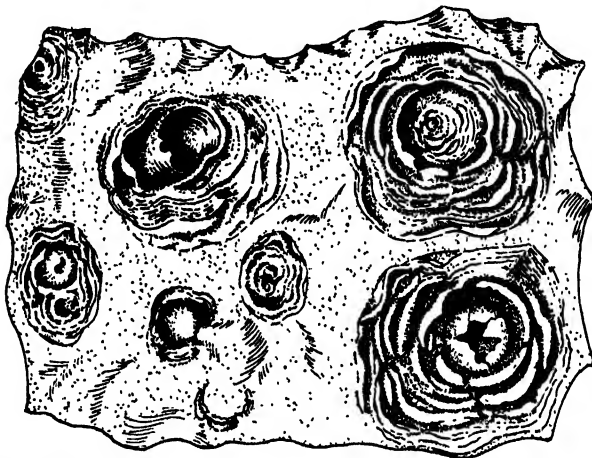


FIG. 92.—Lithophysae (stone-bubbles) in a hand specimen of obsidian, Yellowstone National Park. About natural size.

character of the rock in which it was formed. From its commonly dark color and dull glassy luster it has been named "pseudotachylite." (See the list of further readings.)

VOLCANIC TUFFS

Volcanic dust (and possibly ash up to the size of a pea) when cemented forms tuffs, and may be studied microscopically. Lapilli and bombs form breccia, conglomerate, and agglomerate and are best studied in the field.

Formula.—Fragments either glass and/or minerals and/or rocks.

A. Compositional terms. (Prefix the name of the felsitic rock of similar composition.)

Vitric tuffs—Varieties of vitric tuffs are identified and named as glasses are.

Crystal tuffs } Varieties of lithic and crystal tuffs are named by prefixing the
Lithic tuffs } name of a felsitic igneous rock of similar composition.

Mixed Tuffs—Most tuffs are mixtures of the three above.

B. Characteristics.

1. *Variety.*—The most common tuffs contain all three materials in great variety, i.e., more than three kinds of material.

2. *Form.*—Glass in fibers, bubble walls, and cellular pumice forms, called shards (Figs. 93 and 94). Crystals, some euhedral and corroded. Most rocks and crystal are in angular fragments.

3. *Minerals*.—Mostly from silicic rocks. Usually a variety in the tuff. Rarely make up the whole tuff.

4. *Rocks*.—Any. Mostly the surface lavas, with glasses in which the composition is estimated as in other glasses.

5. *Tuffaceous Sediments*.—Mixed tuff and sediment.



FIG. 93.—Sketch of a thin section of a vitric tuff. The shards are clearly visible by reflected light, even though the glass has been somewhat devitrified. $\times 40$.

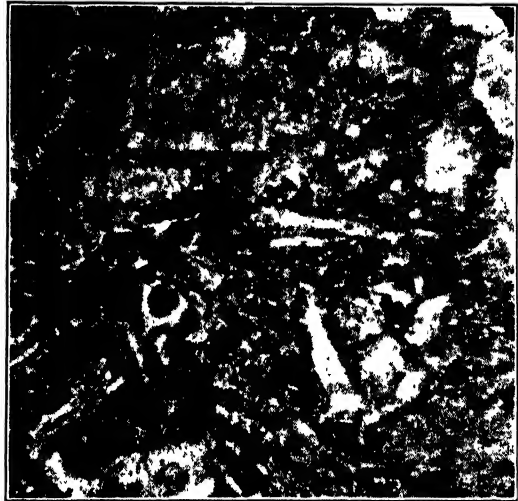


FIG. 94.—Photomicrograph of a thin section of tuff that has been altered. The shards are still recognizable. Plain light. $\times 40$.

6. *Alteration*.—Devitrification, silicification, carbonation and kaolinization may occur with or without complete loss of characteristic forms. Some hydration may occur before crystal structure becomes visible. To make an examination, use plain light with the diaphragm and lower the condenser. Use also reflected light.

Bentonite is a plastic residue from weathering of ash; it swells enormously in water (page 273) and forms a milky suspension; often has a soapy luster.

Regional and contact metamorphism tend to develop sericite and biotite schist from average tuffs. Other varieties may occur.

Jones¹ has discussed the criteria for distinguishing tuffs from other igneous rocks.

7. *Special Names*.

Palagonite	Basic glass in tuff
Porcellanite	Hard, flinty
Häilefinta	Fine textured, acid
Schalstein	Cleaved, basic

NOTE.—Because of the difficulty experienced by prospectors and field geologists in identifying tuffs, the number of tuffs sent to petrographers for microscopic identification is out of all proportion to their abundance in the outcrops.

¹ JONES, F., The petrology (etc.) of Bardon Hill, Geol. Mag., vol. 63, pp. 249–255, 1926.



FIG. 95.—Volcanic bombs of basalt, France. About one-third natural size.

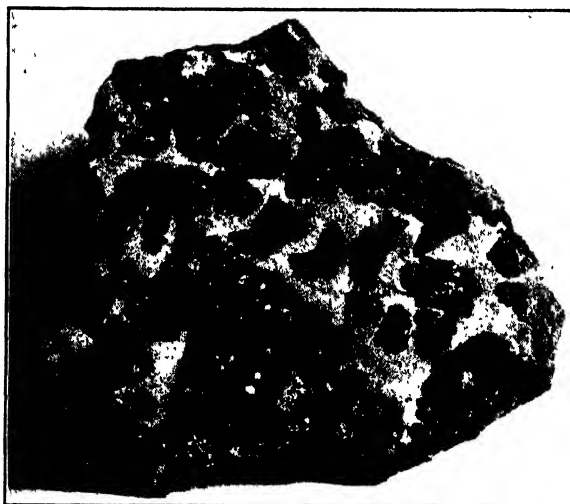


FIG. 96.—Hand specimen of basalt breccia, Pine County, Minnesota. The fragments are amygdaloidal in part, and the cement is largely made up of fine dust from a basaltic rock. About one-half natural size.

IGNEOUS BRECCIAS

Volcanic fragmental rocks coarser than ash and tuff include *lapilli*, the size of nuts, and *bombs* (Fig. 95) the size of apples or larger. *Cinders* is a term for a mixture of ash and lapilli. A rock composed of any of these coarse materials cemented with either ash or lava, or vein

materials, is a volcanic breccia. Cemented material of mixed sizes is *agglomerate*.

1. *Eruptive breccias* may contain interspersed fragments of sedimentary and crystalline rocks from the walls of the conduit. They characteristically form from viscous acidic magmas but basalt breccias are not rare (Fig. 96). Breccias are irregularly stratified and many are



FIG. 97.—Outcrop of a contact breccia, Nigadoo, Minnesota. A diabase intrudes massive anorthite.

interbedded with felsitic flows, or if derived from submarine volcanoes interbedded with sedimentary rocks. The majority were formed by the explosion of material in the crater, but some breccias seem to have flowed as an incandescent cloud for great distances from the crater.

2. *Flow breccias* consist of angular, brecciated fragments in a basaltic or felsitic matrix, and represent pyroclastic material thrown out and scattered over an unconsolidated flow. They may also result from the collapse of thin surface crusts of solidified material on a flow.

3. *Contact breccias* are due to the shattering of the wall rocks around laccolithic and other igneous masses. The breccia may consist of fragments of both the wall rock and the intrusive. The outside of the igneous mass may crystallize and later be brecciated before the entire mass solidifies. The intrusive serves as a cement (see Fig. 97).

See the list of further readings.

ABUNDANCE AND OCCURRENCE OF ROCK TYPES

TABLE IV.—APPROXIMATE QUANTITATIVE DISTRIBUTION OF IGNEOUS ROCKS IN THE UNITED STATES
(After Daly)

Forms of rock mass	Rock	Square miles	Per cent
Batholiths.....	{ Granites	3,836	23
	{ Granodiorites	2,040	12
Sills and dikes.....	{ Gabbro	274	2
	{ Diabase	268	2—
Extrusive flows.....	{ Rhyolite	2,147	13—
	{ Andesite	4,245	25
	{ Basalt	3,209	19
All others.....		713	4+
		16,732	100

About 95 per cent of all sills are diabase.

Typical laccoliths have trachite porphyry but vary considerably.

All the "alkalic" rocks together—phonolites, nephelite syenites, monzonites, latites, missourites, fergusites, etc.—make less than 1 per cent of the total area and no doubt less than one-half of 1 per cent of the volume of the igneous masses; probably less than 0.1 per cent.

The following have small *total known* area:

	Square Miles
Essexite.....	50
Shonkinito	20
Theralite.....	20
Ijolite	5
Missourite..	1
Fergusite	1
Bekinkinito ...	1

Others occupy still smaller areas. Contrast these minute volumes with the fact that about three-fourths of the literature of petrography deals with the rare rocks.

The granite clan is highly predominant in intrusives but somewhat subordinate among extrusives. In the pre-Cambrian areas granites probably constitute much more than three-fourths of all exposures.

The gabbro and diorite clans are very subordinate in intrusives but distinctly predominant in extrusives.

Quantitatively the rocks of the earth crust are *granite* and *basalt*. The batholiths are granite and granodiorite and the flows are basalt and andesite. This is very significant in geologic theory.

Several petrographers emphasize the relation between the type of igneous rock formed and the nature of earth and magma movements. Granites characterize the great shield areas of the earth, whereas granodiorites and andesites are characteristic of the earth's "hinge zone," the belt of active volcanoes and present-day earthquakes (see page 131).

"Volcanoes" in the sense of conical mountains furnish probably less than 0.1 per cent of our igneous rocks. "Fissure eruptions" are perhaps less numerous but are of much greater volume and extent. The fissure eruptions develop no mountains but rather a plain sloping one or two degrees. Only one such eruption is recorded in historic times—that at Laki, Iceland, in 1783, which spread a sheet 100 feet thick over hundreds of square miles (Fig. 9). Compare this with the historic flows of Vesuvius, perhaps a mile long and a few feet wide.

The great fields of lava include the Archean of the Lake Superior and other regions, the Keweenaw of the Lake Superior region, the Columbia River and Snake River Series, the Purcell series in British Columbia, the Appalachian Tertiary series, the Deccan flows in India, the British Tertiary series, Icelandic flows, and Mexican flows.

DIKE ROCKS IN CLASSIFICATION

Dikes of different compositions occur near large igneous masses and some are called "complementary," because if taken in the proper propor-

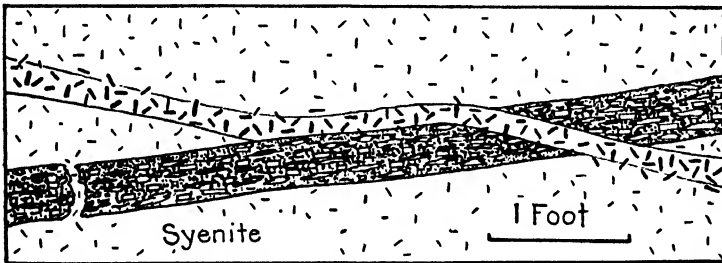


FIG. 98.—Dikes in syenite at Snowbank Lake, Minnesota. The lamprophyre seems to have been intruded before the syenite was entirely crystallized, for some syenite, continuous with the main mass, fills a small gap in the broken dike. The pegmatite seems to be a late residual magma from the same syenite not far away; its walls are not sharp even here in the area sketched.

tion they may yield a composition like that of the large mass they accompany. Although this in no way proves a splitting of the magma into two extreme liquids, one must admit that it strongly suggests such a splitting, and the study of dikes for many years has been based largely on this notion. The light or silicic dike rocks are leucophyrs or aplites; the darker ones are lamprophyrs. Dikes that seem to have split are called diaschistic, signifying "split in two"; others are aschistic, or "not split." Although many of the occurrences may have basic dikes early and silicic dikes late, in accord with the theory of differentiation by crystal settling, other regions show basic dikes of a late magmatic stage. The two may be almost simultaneous.

This idea of splitting, however, is theoretical and unsafe as a basis of classification. The more exact basis for separation of this group of rocks

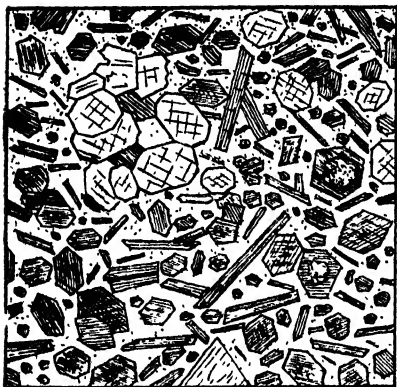


FIG. 99.—Sketch of a thin section of monchiquite. The tendency to be sugary-grained is evident in spite of some analcite as a groundmass. $\times 25$.

is the combination of certain mineral and textural peculiarities. If the dike rocks have a "sugary" texture, they may deserve distinct names, so that they will not be confused with granitoid or felsitic rocks of the more common sorts. The following three groups will serve to illustrate the usage. Most lamprophyrs are intrusive into granite or syenite at such a stage of magmatic history that they are easily deformed and intruded by the late residual magma of the granite (see Figs. 98 and 130).

The term lamprophyr for the darker diaschistic dikes is much used.

In 1874 Gümbel gave the term to certain dark dikes unusually rich in mica. At present there is much diversity in usage, but usually the term

TABLE V.—THE RELATIONS OF DIASCHISTIC DIKES

Expected leucophyr	Large intrusives	Expected lamprophyr
Aplite Pegmatite	{ GRANITE SYENITE }	Minette Vogesite Kersanite Spessartite Durbachite
Bostonite Tinguaite Paisanite	{ NEPHELITE SYENITE DIORITE MONZONITE }	Minette Vogesite Kersanite Camptonite Monchiquite Fourchite Alnoite Ouachitite
Malchite.	{ GABBRO DIABASE PERIDOTITE }	Camptonite ¹ Monchiquite Alnoite Issite Odinite Iron ores

¹ A supposed "gabbro aplite," beerbachite, in Odenwald, proved to be a hornfels. See Geol. Mag., vol. 68, pp. 506-521, 1931

signifies rocks that are dark, basic, mostly of sugary texture (Fig. 99), and mostly occurring in dikes. They have abundant mafic minerals and commonly considerable potash, and they alter readily to calcite (Figs. 100 and 101). They are not so coarse as average granite but may have

phenocrysts and may be glassy, at least in part. Those who base their rock classification on occurrence commonly find it best to classify lampro-

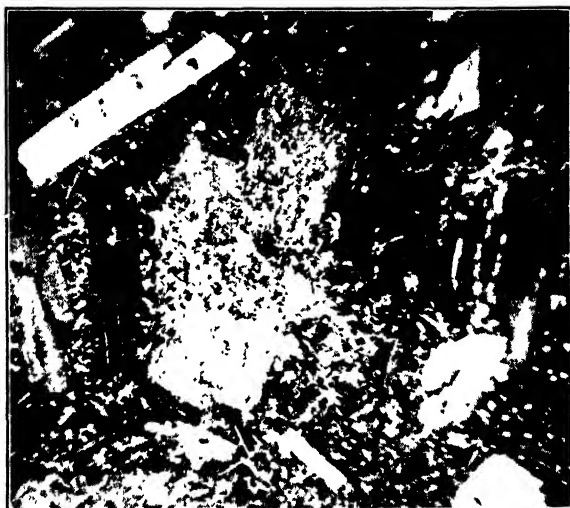


FIG. 100.—Photomicrograph of a lamprophyre with a large phenocryst (gray) nearly all of which is replaced by granular calcite, although the associated minerals are fresh. Plain light. $\times 60$.

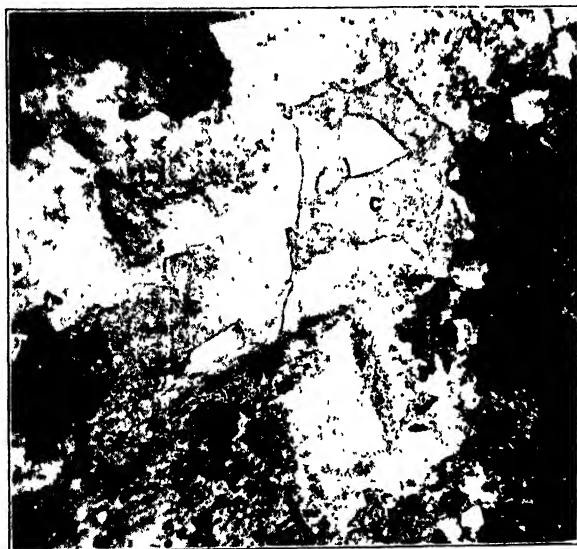


FIG. 101.—Photomicrograph of syenite in which the calcite (C) seems to be primary though later than feldspar. $\times 50$.

phyrs rather differently from other igneous rocks because the feldspars are subordinate and altered and the potash is present largely in the micas. Probably the most useful system of naming these rocks would be to

prefix the name of the chief mafic mineral to the term lamprophyr, but a long list of special names has been applied to minor varieties. (See the list of further readings.)

TABLE VI.—TENTATIVE CLASSIFICATION OF LAMPROPHYRS

Group 1. Hornblende dominant (or some related amphibole)							
	With augite					With biotite	
	With feldspathoids and feldspars		With no feldspathoid			With quartz	No quartz
			+ plagioclase		+ orthoclase		
	+ olivine	— olivine	+ olivine	— olivine	Vogesite	Malchite Gladkaite Lucite	
Granitoid, often sugary			Spessartite				
Porphyritic....	Giumarite Epichellite	Monhaldeite Heptorite Amphibole- fourchite		Camptonite			Kvellite Orbite
Glassy.....	Heumite						
Group 2. Biotite dominant							
	With melilite — feldspar — olivine	With augite and plagioclase With no feldspathoids		With hornblende and orthoclase With no feldspathoids			
		+ olivine	— olivine	+ quartz	+ olivine		
Granitoid, often sugary		Kersanite		Aschaffite	Minette Durbachite		
Porphyritic..	Alnöite	Cascadite	Ouachitite				
Glassy.....	Bergalite						
Group 3. Augite dominant (or some related pyroxene)							
	With olivine			With no olivine			
	+ feldspathoid		— feld- spathoid	+ biotite		± hornblende	
	+ feldspar	— feldspar		+ feldspar	— feldspar	+ feld- spathoid ± feldspar	— feld- spathoid + feldspar
Granitoid, often sugary....	Polzenite Bekinkinite					Tamaräite	
Porphyritic.....	Monchiquite (Some varieties with biotite and amphibole)		Garewaite Tjoseite	Topsallite Ouachitite		Fourchite	Odinite

SPECIFIC GRAVITIES OF IGNEOUS ROCKS

A compilation showing the number of rocks in each family that have a certain specific gravity shows the dominance of the following:¹

Granites.....	2.66	Quartz porphyry.....	2.62
Granodiorite.....	2.71		
Quartz diorite.....	2.80		
Diorite.....	2.87		
Gabbro.....	3.00	Basalt.....	2.85
Pyroxenite.....	3.25		
Peridotite.....	3.23		
Iron ores.....	4.13		
Average igneous rock of the crust..	2.761		

MINERAL PROPORTIONS IN IGNEOUS CLANS

Through the series of clans the changes in abundance of the important mineral groups may be roughly estimated as follows, remembering that wide variations occur in each clan:

Rock family	Mafic, per cent		Felsic, per cent	
Granite.....	Biotite,	10	{ Orthoclase, 30 Albite, 30 Quartz, 30 }	90
Syenite.....	Hornblende,	20	{ Orthoclase, 35 Albite, 40 Others, 5 }	80
Shonkinite.....		60		40
Nephelite syenite.....	{ Aegirite Biotite }	{ 15 to 20 }	{ Orthoclase, 25 Albite, 35 Nephelite, 25 }	{ 85 to 80 }
Granodiorite.....		15		85
Monzonite.....	{ Hornblende Biotite }	{ 15 to 40 }	{ Orthoclase, 35 Andesine, 40 Quartz, 10 }	{ 85 to 60 }
Quartz diorite.....		25		75
Diorite.....	{ Hornblende (Augite) (Magnetite) }	{ 35 to 45 }	{ Orthoclase, 10 Andesine, 50 Quartz, 4 }	{ 65 to 55 }
Gabbro.....	{ Augite Olivine Magnetite }	{ 45 to 50 }	Labradorite	{ 55 to 50 }
Basanite, Feldspathoid basalt..		45		55
Pyroxenite, etc.....	{ Pyroxene Olivine Ores }	{ 95 to 98 }		{ 5 to 2 }

This summary emphasizes the dominance of felsic minerals in igneous rocks as a whole (Fig. 102).

The changing nature of the varietal and accessory and secondary minerals, though not conspicuous in tracing through the clans, is very

¹ For further detail see HOLMES, ARTHUR, "Petrographic Methods and Calculations," pp. 23-25, London, 1923.

striking when the granites are contrasted with the gabbro or following clans. Zircon, titanite and fluorite give way to chromite, perovskite and

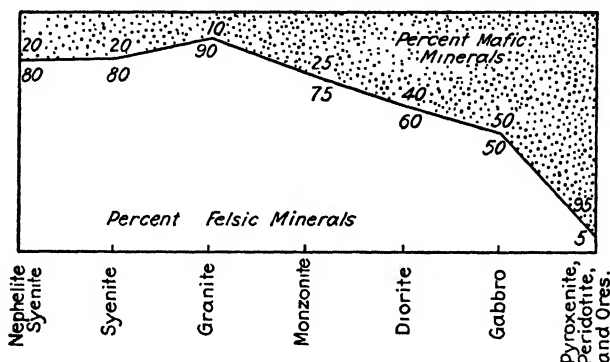


FIG. 102.—The curve of mineral proportions through the principal igneous rock clans.

spinel. Sericite, kaolinite, and chlorite give way to leucoxene, serpentine, and talc.

CHEMICAL COMPOSITION OF IGNEOUS ROCKS

In the earth's crust to a depth of 10 miles, it is estimated that more than 98 per cent of the rocks consists of the first nine oxides in the follow-

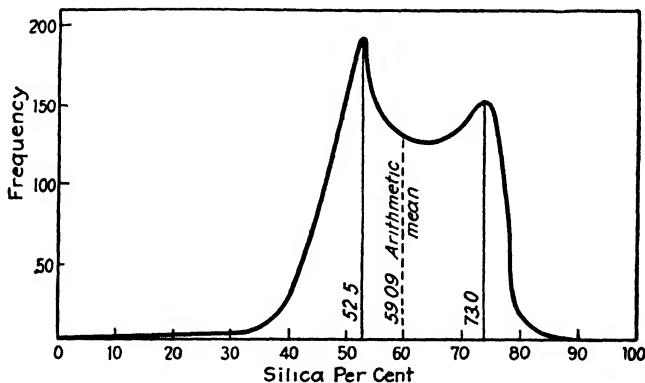


FIG. 103.—Frequency-distribution of silica in Washington's collection of analyses of igneous rocks, U. S. Geol. Survey Professional Paper 99. (After Richardson and Sneesby, *Mineralogical Mag.*, vol. 19, p. 306, 1922.)

ing list. These few constituents clearly determine the most important characters of rocks. Several thousand analyses have been made on igneous samples, but they show a wide range in quality. Errors of one per cent or more may be expected in the abundant constituents of many reported analyses. For the purpose of this summary, however, the results have been averaged by groups to eliminate individual errors, and the errors due to the method of averaging are probably small.

TABLE VII.—AVERAGE PERCENTAGE COMPOSITION BY FAMILIES¹

	Granite	Pegmatite ¹	Syenite	Nephelite syenite	Monzonite	Granodiorite	Diorite	Gabbro	Pyroxenite	Peridotite	Ores ¹
SiO ₂	69.92	71.80	60.19	54.63	55.25	65.10	56.77	48.24	51.29	41.09	10.44
Al ₂ O ₃	14.78	15.50	16.28	19.89	16.53	15.82	16.67	17.88	3.52	4.80	4.78
Fe ₂ O ₃	1.62	0.60	2.74	3.37	3.03	1.64	3.16	3.16	1.82	3.96	31.05
FeO.....	1.67	0.65	3.28	2.20	4.37	2.66	4.40	5.95	6.00	7.12	24.67
MgO.....	0.97	0.30	2.49	0.87	4.20	2.17	4.17	7.51	21.06	32.25	3.77
CaO.....	2.15	1.10	4.30	2.51	7.19	4.66	6.74	10.99	13.88	4.42	5.08
Na ₂ O.....	3.28	3.75	3.98	8.26	3.48	3.82	3.39	2.55	0.30	0.49	0.17
K ₂ O.....	4.07	5.30	4.49	5.46	4.11	2.29	2.12	0.89	0.16	0.96	0.11
H ₂ O.....	0.78	0.50	1.16	1.35	0.66	1.09	1.36	1.45	1.20	3.53	0.62
P ₂ O ₅	0.24	0.15	0.28	0.25	0.43	0.16	0.25	0.28	0.06	0.12	2.02
TiO ₂	0.39	0.15	0.67	0.86	0.60	0.54	0.84	0.97	0.58	1.16	18.01
MnO.....	0.13	0.15	0.14	0.35	0.15	0.05	0.13	0.13	0.13	0.10	0.23
Ratio $\frac{Na_2O}{K_2O}$	0.8	0.7	0.9	1.5	0.8	1.7	1.6	2.9			

¹ The analyses are mostly from Daly, "Igneous Rocks and Their Origin." The pegmatite is an average from U. S. Geol. Survey Prof. Paper 99, and the ores are the average of 24 in J. P. Iddings, "Igneous Rocks," vol. 2, pp. 340-341.

Notes on Averages of Analyses of Clans.—There is more variation in one family than from one to the next, and still the average gives a useful standard for comparison.

The Al_2O_3 shows slight range—14 to 20 per cent—except in the three basic clans.

Alkalies are maximum in nephelite syenites.

Lime is maximum in pyroxenites.

Magnesia is maximum in peridotites.

Iron is maximum in ores.

The alkalies and silica largely determine the nature of the rock (see next section).

} The three segregations in gabbro
magmas.

Examples of individual analyses are found in many of the larger works; and, in Professional Paper 99 of the United States Geological Survey, H. S. Washington has given a compilation of analyses published up to 1913. From that work the distribution of silica content has been plotted in a curve (Fig. 103). It is noteworthy that few rocks have less than 30 or more than 80 per cent of silica. Similarly, few have less than 2 or more than 5 per cent of soda. The normal range of igneous rocks is somewhat restricted.

AVERAGE IGNEOUS ROCK OF THE EARTH'S CRUST¹

	Per cent		Per cent
SiO_2	59.12	S.....	0.052
Al_2O_3	15.34	F.....	0.030
Fe_2O_3	3.08	$(\text{Ce}, \text{Y})_2\text{O}_3$	0.020
FeO	3.80	Cr_2O_3	0.055
MgO	3.49	V_2O_5	0.028
CaO	5.08	MnO	0.124
Na_2O	3.84	NiO	0.025
K_2O	3.13	BaO	0.055
$\text{H}_2\text{O} +$	1.15	SrO	0.022
CO_2	0.102	Li_2O	0.007
TiO_2	1.050	Cu	0.010
ZrO_2	0.039	Zn	0.004
P_2O_5	0.299	Pb	0.002
Cl	0.048		
			100.000

¹ From U. S. Geol. Survey Prof. Paper 127, p. 16.

This analysis resembles that of quartz-monzonites.

It is noteworthy that oxygen is closely combined with all the abundant elements, and a study of the space lattice indicates that the arrangement of the oxygen probably controls the structure. The "oxygen volume" is said to be more than 90 per cent in the lithosphere.¹

IMPORTANCE OF ALKALIES IN IGNEOUS ROCKS

If a representative large series of igneous rocks is plotted with silica and alkali contents as coordinates, it will be noted that there is a suggestion of two series diverging in alkalies as silica decreases, (Fig. 104).

¹ GOLDSCHMIDT, V. M., Ueber die Raumerfüllung der Atome in Kristallen und ueber das Wesen der Lithosphäre, Neues Jahrb. Fest., pp. 1119–1130, Otto Mügge, 1928.

Although intermediate rocks are known, there are very few. This perhaps justifies the emphasis placed by some petrographers on the alkalis, though similar divisions might result from the plotting of other

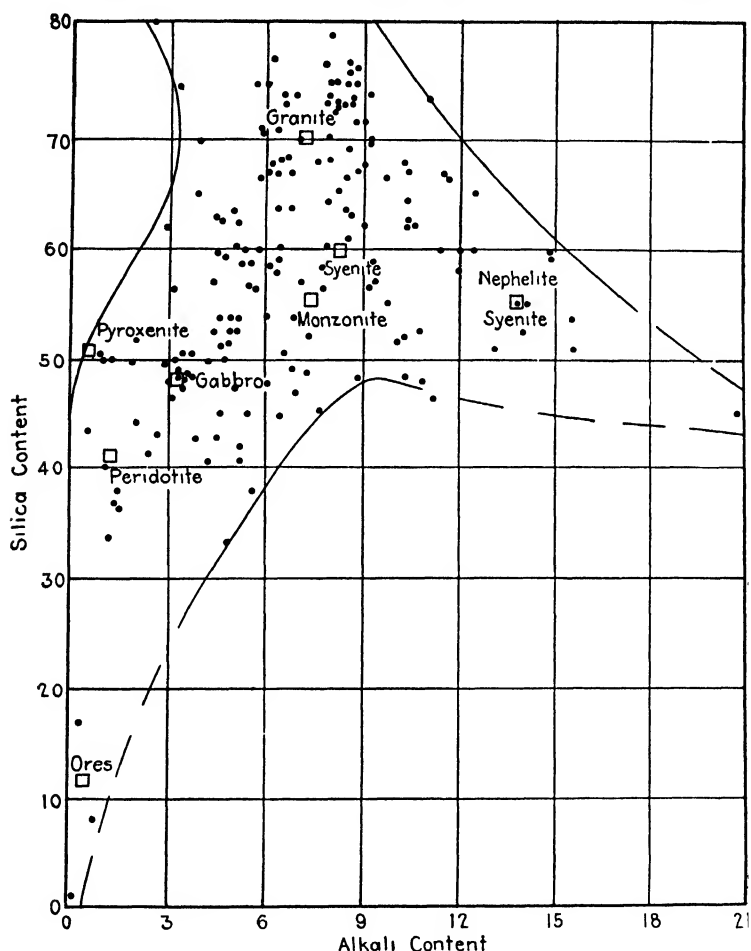


FIG. 104.—The igneous rock "field" plotted on coordinates of silica and alkali percentages. Data from Washington's tables. Each spot is a representative of an arbitrarily selected group of analyses. Squares for the main rock clans are plotted from Daly's average analyses.

constituents. Rocks with about 50 per cent silica can be divided into calcic and magnesian types as clearly as into alkalic and subalkalic types.

The alkalic rocks as a distinguishable series are not always sharply defined. It is generally agreed that feldspathoids and alkalic pyroxene indicate an alkalic nature, but not so generally agreed that abundant orthoclase and albite, the alkalic feldspars, make a rock alkalic. It must be admitted, however, that even a granite may vary widely in the percentage of alkali present. A granite with abundant quartz and hornblende and a good deal of calcic oligoclase or andesine can not have as much

alkali as a granite with only orthoclase, albite, mica, and a small amount of quartz. Even this latter rock, however, has no mineral that is alkalic in the sense of having alkali in greater ratio to silica than 1:6 (see pages 48 and 49). Many alkalic granites contain such minerals as arfvedsonite or acmite, in which the alkali is in proportion to silica as 1:4. These are to be classed as more alkalic in general than those with common mica and alkalic feldspars.

In fact it is probably true that the emphasis given in recent years to alkalic rocks has been due to the occurrence of strikingly peculiar minerals such as acmite and arfvedsonite. These at once attract the attention of a petrographer who deals largely with the normal series of rocks from granite to gabbro and peridotite. There are as many special and locality names for nephelite syenites as there are for much more abundant special phases of granites. (See the list of further readings.)

IGNEOUS ROCK SERIES¹

If all igneous rocks are considered together, it is clear that there is no break in the series, but a complete gradation from each type to all the others (see Figs. 102 and 104).

There are, furthermore, broad regions in which a wide variety of rocks have some peculiarity in common, strongly suggesting a genetic relation. Such areas in which the rocks present a certain degree of consanguinity but differ from those in adjoining areas are called "petrographic provinces."² Harker³ has suggested that the whole earth may be divided into two petrographic branches—the "Atlantic," in which alkali feldspars are abundant in the silicic and intermediate rocks and occur even in the less silicic rocks, and the "Pacific," in which alkali feldspars are not abundant except in the more silicic rocks. Briefly, but less accurately, the Atlantic rocks constitute an alkalic group and the Pacific rocks a subalkalic or calci-alkalic group. As these terms have been used by various writers the Atlantic has come to suggest sodic rocks, the Pacific to represent calci-alkalic rocks, and a third group has been proposed, the Mediterranean, for potassic rocks.⁴ Although these greater provinces may show real differences, the exceptional occurrences are numerous. There is not only one Pacific province but several. The type of rock series in a province should be correlated not with its geographic position but rather with its tectonic setting, which may determine the type of differentiation. Sunken areas are mostly Pacific. Orogenic areas show Pacific and Mediterranean series. Borders of orogenic belts are Atlantic and Mediterranean series. Farther out in the shield areas the rocks are mostly Atlantic⁵ (Fig. 105). Roughly, the alkalic series appear where the

¹ The textural series so commonly found near the chilled borders of igneous masses have already been mentioned and are not discussed here.

² A term proposed by Judd, *Geol. Soc. London, Quart. Jour.*, vol. 32, p. 292, 1876; vol. 42, p. 54, 1886. H. S. Washington prefers "comagmatic regions."

³ HARKER, ALFRED, "Natural History of Igneous Rocks," pp. 88-104, The Macmillan Company, 1909.

⁴ NIGGLI, PAUL, "Gesteins- und Mineral-provinzen," p. 96, Borntraeger, 1923; also an article in *Schweiz Min. u. Pet. Mitt.*, vol. 2, pp. 169-275, 1923.

⁵ BURRI, C., Differentiation in post-Mesozoic series, *Schweiz Min. u. Pet. Mitt.*, vol. 7. No. 2, pp. 254-310, 1927.

crust has been subject to slow vertical movements, in contrast to the calc-alkalic series, which are associated with compression and folding.

The terms Atlantic, Pacific and Mediterranean are best replaced by chemical terms, and there are not only these three suites but also some that are ferromagnesian and others with still other peculiarities, and there are gradations from each to the others.

Groups of rocks that show consanguineous characters and that appear to be genetically related are called by Tyrrell¹ kindreds, series, suites, tribes or clans. Kindreds

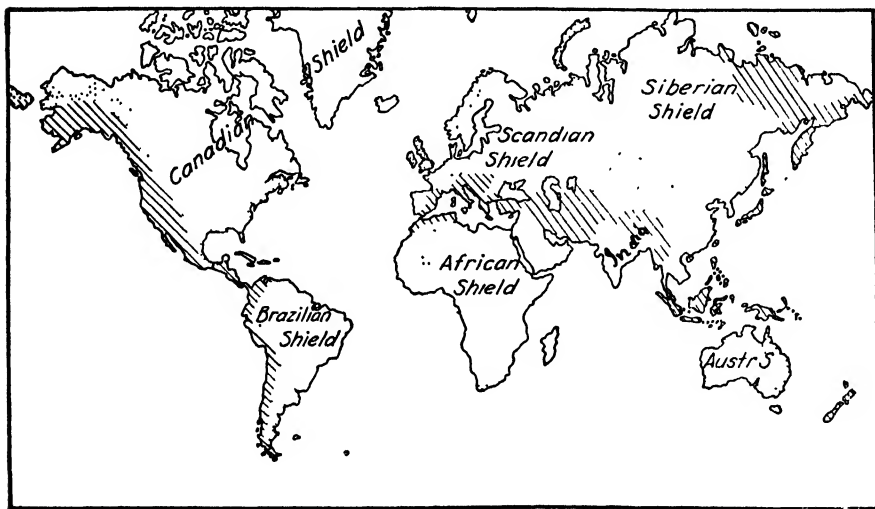


FIG. 105.—Map of the world showing the distribution of "shields," "bow areas" (stippled) and "hinge zone" (line shading). (Modified from L. Kober.)

cut across formal classifications, and rocks of certain families may recur in several different kindreds. The kindreds supposedly established are: (1) Spilitic, characterized by abundant soda; (2) Arctic, characterized by plateau basalt; (3) the granite-diorite kindred, equivalent to the rocks of the calcic province; (4) the anorthosite-charnockite kindred; (5) the richly alkalie rock kindred; (6) Mediterranean or potash-rich kindred.

These broad compilations, however, are not the most definite available evidence that a series of rocks in different clans may be genetically related. In many places lavas of different compositions are extruded from a single crater, and best of all there are large outcrops in which a visible gradation can be traced from one rock type to another. Examples of these gradations are of considerable importance as a basis of igneous theory.²

Rock Series in Single Masses.—1. *Mountain batholiths* show gradations from granite or granodiorite, which makes up the large bulk of the

¹ TYRRELL, G. W., "Principles of Petrology," pp. 136-142, E. P. Dutton & Company, Inc., 1926.

² Daly has compiled a large number of examples in "Igneous Rocks and Their Origin," pp. 230-236 and Appendix B.

masses, through diorite and gabbro to peridotite. Pegmatites are abundant near the contacts. These gradations may be considered the normal and most common batholithic series. In a few masses the granite grades into alkalic phases and nephelite syenites. In some batholiths of the Canadian shield granites grade through syenite and shonkinite to hornblendite and iron ores and locally show, parallel to the walls, vertical flow layers of several differentiates.¹ Several masses of granite without any wide variation in silica show sodic feldspar at the border and become progressively potassic toward the center.² Those granites that have been contaminated by clay rocks from their walls may show a series grading to cordierite or sillimanite granite. Finally, there are very locally near some inclusions in granites certain modified igneous rocks which contain constituents other than those derived from the inclusions,³ apparently segregated from the magma.

2. *Plateau batholiths* show similar series from gabbro to granite.

3. *Stocks* show rock series much like those of batholiths. There is perhaps a tendency for the smaller masses related to batholiths to be less siliceous and more alkalic than the larger masses. Monzonites and syenites grade on one hand into granite and on the other to diorite, malignite, shonkinite, and hornblendite. Normally the shonkinite is a border phase of granite, but in some stocks the outer border is more siliceous rather than more basic. Even in these masses the granite seems to be a late phase of the magma.

4. *Pipes* or *plugs* may show a series of rocks grading into each other concentrically and with vertical banding. Although most plugs are made up of common aphanites, some of them have alkalic rocks.

Stocks and plugs (and the lopoliths referred to below) show series related to their walls and included fragments. Cordilleran intrusives are notably more calcic, richer in dark minerals like diopside, near limestones than near other wall rocks. At Sudbury, Ontario, inclusions that appear to have been quartzite have contaminated the main norite magma to form conspicuous series.

5. *The type of lopolith* is partly defined by the remarkable series of rocks developed—gabbros containing peridotite, troctolite and basal segregations, and an upper zone of granite. There may be norite with sulphides at the bottom, or banded chromite and platinum segregations. Alkalic syenites came later, possibly a whole series of them in banded segregations.

The lopoliths differ from batholiths and stocks in having heavy rocks near the floor and light rocks near the roof, whereas in batholiths as a rule

¹ See the U. S. Geol. Survey Folios covering Cordilleran regions; Geol. Survey of Canada Mem. 6; and Minnesota Geol. Survey Bull. 21, pp. 27-76.

² GROUT, F. F., The Saganaga granite of Minnesota-Ontario, Jour. Geology, vol. 37, pp. 562-591, 1929.

³ GROUT, F. F., Anorthosite and granite as differentiates of a diabase sill at Pigeon Point, Minnesota, Geol. Soc. America Bull., vol. 39, pp. 572-573, 1928.

the heavy rocks are near the borders and the floors are not exposed. In the logical development of igneous theory these masses and the following, in which there is a differentiation and in which the whole series from roof to floor and from side to side is exposed for study, are especially valuable.

6. *Laccoliths* have furnished some remarkably exposed series. A few are worthy of individual description. Shonkin Sag, Montana, is almost diagrammatic in its exposure (Fig. 106). It has a border of leucite

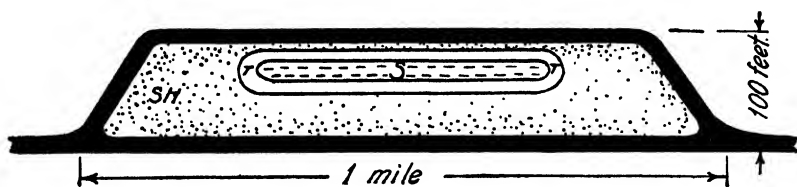


FIG. 106.—The laccolith of Shonkin Sag, Montana. (After Weed and Pirsson.) Vertical scale greatly exaggerated. Leucite basalt porphyry border and continuing sheet (black); shonkinite (SH); transition rock (T); syenite (S).

basalt, probably chilled before differentiation, and a core of shonkinite below grading to syenite above.¹ The Borolan mass in Scotland is highly alkalic and calcic and may have been modified by assimilation of limestones. It grades from quartz syenite to borolanite, with several odd nephelite, leucite and garnet rocks, possibly from more than one intruded magma. The Raana mass in Norway is said to have an olivine norite segregated both above and below a quartz norite lens in a schist² (Fig. 107).

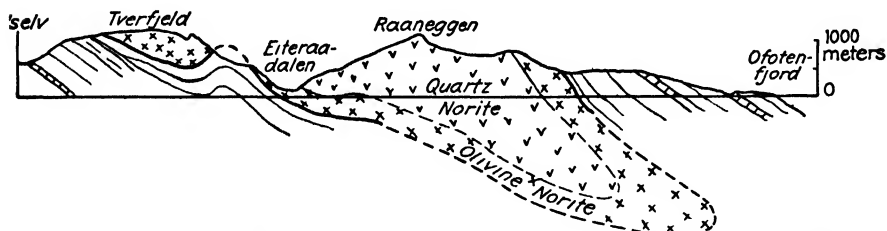


FIG. 107.—Section through the Raana, Norway, intrusive. (After Foslie, *Jour. Geology*, vol. 29, p. 707, 1921.) Vertical and horizontal scales the same.

An intrusive mass at Carrock Fell underlies and includes fragments of a series of lavas. The lavas are now somewhat steeply tilted, and it seems likely that the intrusive underlying the flows was tilted up with the flows. If so, the intrusive was probably a laccolith, and its top was next

¹ PIRSSON, L. V., *Igneous rocks of the Highwood Mountains*, U. S. Geol. Survey Bull. 237. The outer border was supposed by Pirsson to be an early differentiate, but it is very much like the undifferentiated small sheets. F. F. OSBORNE and E. J. ROBERTS, *Differentiation in the Shonkin Sag laccolith, Montana*, *Am. Jour. Sci.*, vol. 22, pp. 331–353, 1931.

² FOSLIE, STEINAR, *Field observations in northern Norway*, *Jour. Geology*, vol. 29, p. 705, 1921.

to the flows. The intrusive, however, was markedly differentiated in a series that varies chiefly in proportions of the minerals present. It is

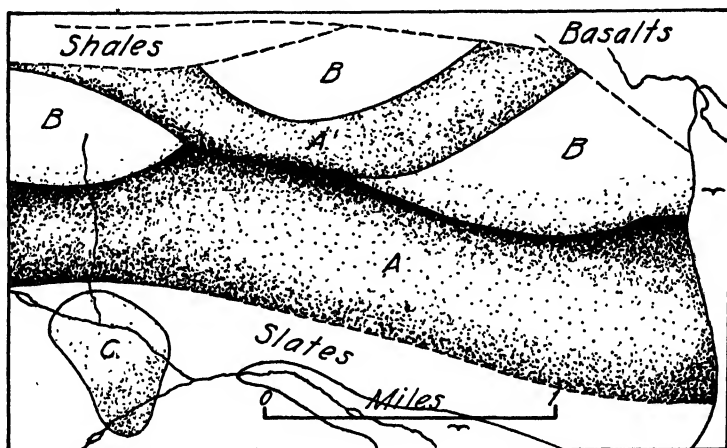


FIG. 108.—Sketch map of Carrock Fell, Cumberland. (After Harker.) The igneous rocks show three kinds of series. The gabbro (A) is enriched in ilmenite as indicated by shading. The granophyr (B) is most basic along the southern border where it intrudes the gabbro and has assimilated some of it. The greisen of Grainsgill (C) is a modified pegmatite.

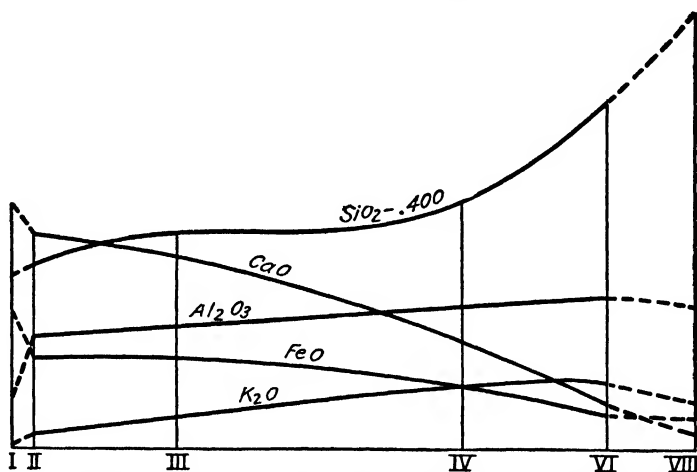


FIG. 109.—Diagram of the molecular ratios of some of the constituents of the rock series at Magnet Cove, Arkansas. Abscissas represent distances from the origin at II. (After Washington, *Jour. Geology*, vol. 9, p. 651, 1901.)

well exposed in a cross section cut by erosion of the tilted mass (Fig. 108). If the suggested form is correct, ilmenite is greatly concentrated at both the roof and floor, and gabbro with some granophyric phases increases toward the center.¹

¹ HARKER, ALFRED, Carrock Fell, *Geol. Soc. London, Quart. Jour.*, vol. 50, p. 311, 1894.

The intrusive peridotite in North Carolina, that was supposed to have corundum differentiated to the sides and roof, is now believed to be a complex of pegmatites desilicated by peridotite.¹

The intrusive complex at Magnet Cove, Arkansas, a stock or laccolith, exhibits a series of alkalic rocks, shown in Fig. 109.²

7. *Sills*. Several sills 500 to 1,000 feet thick have near the bottom a zone of peridotite or olivine-rich rock; above this is an olivine diabase grading into quartz diabase and granophyr, becoming more silicic and alkalic above. Some have chilled margins at the top. Some sills may have the upper half filled with coarse phenocrysts of plagioclase apparently floated up from the lower half.



FIG. 110.—Intrusive relations in the differentiates of the Duluth gabbro. The light gray rock is anorthosite near the top of the chamber. The darker stringers are gabbro and red rock (aplite) derived from the same magma. The contacts, especially those of gabbro and red rock, are gradational.

The Lugar sill in Scotland has been described as having a series of zoned differentiates.³ Later study indicates successive intrusions in which some differentiation occurred, but it is doubtful if all the series, including teschenites, picrite, theralite and lugarite, separated in the present chamber. It is not at all uncommon for late liquid fractions of a magma to intrude early solidified fractions as dikes or stringers

¹ GORDON, S. G., Desilicated granitic pegmatites, Philadelphia Acad. Nat. Sci. Proc., 1921, pp. 169–192; and MARGARET C. COBB, Desilicated genesis of commercial corundum; Pan-Am. Geologist, vol. 43, p. 257, 1925.

² WASHINGTON, H. S., The foyaite-ijolite series of Magnet Cove, Jour. Geology, vol. 9, p. 646, 1901.

³ TYRRELL, G. W., The picrite-teschenite sill of Lugar, Geol. Soc. London, Quart. Jour., vol. 72, pp. 84–131, 1917; and Geol. Mag., vol. 9, p. 75, 1912.

(Fig. 110), and it is one of the characteristics of aplites, lamprophyrs and pegmatites that the walls may be sharp in some places but gradational in others.

There is said to be a sill in Formosa with sodic minerals and olivine concentrated in the center but much less of them above and below.¹

Note that there are really very few examples (Raana, Norway, Formosa, and possibly Carrock Fell) in which it seems clear that a heavy early differentiate forms above as well as below the late differentiate of the same mass. In general the arrangement is gravitative. See the criteria tabulated for distinguishing two intrusives from differentiated masses, page 259.

8. *Flows.* A few large lava flows show a slight difference, chiefly a concentration of phenocrysts in the upper or lower parts.

9. *Dikes.* A few large dikes show a slight difference in composition from side to center. Most of the dikes that show rock series are pegmatites, many of which grade from granite to silexite. The gradation may be along the dike or it may be to a central zone of quartz between borders of granite. A more surprising variation in pegmatite is that from gabbro borders to granite center.

It should be added that the variation in single masses is not great except in those bodies that cool slowly enough to have granitoid textures. Aphanites rarely show connected series.

Rock Series in Discontinuous Exposures. *Petrographic Provinces.*—The characteristics of certain regions prove very noteworthy.² Southern Norway is sodic; Montana and Italy are potassic; eastern Spain is rich in TiO_2 ; Pantelleria is rich in iron; Japan is rich in calcium and aluminum. The genetic connection of the several rocks that exhibit these peculiar chemical features in a province will probably never be proved, but the peculiarities are suggestive.

Complementary Dikes.—Commonly associated with intrusive bodies such as stocks are groups of dikes. These range from the same composition as the stock to both silicic and basic extremes. The silicic and basic dikes may be complementary (page 121), strongly suggesting their genetic connection with each other and with the main intrusive.

Flows from a Single Vent.—The succession of flows from a crater may range from rhyolite to basalt, or from alkalic to calcic types. In such a series there may be flows of intermediate composition, but in some volcanic piles, rhyolite and basalt alternate without intermediate rocks. No series of flows can be set up as a very definite standard.³ The rocks of the Katmai volcanic region form a very complete series from olivine

¹ ICHIMURA, TAKESHI, personal communication.

² PIRSSON, L. V., The petrographic province of central Montana, *Am. Jour. Sci.*, 4th ser., vol. 20, pp. 35–49, 1905.

³ DALY, R. A., "Igneous Rocks and Their Origin," Appendix B, 1914, contains a very large compilation of observed sequences.

basalt to a siliceous soda rhyolite,¹ but this is not a series in regular time sequence.

The Trend and Sequence of Rock Series.—One of the most common and normal series in the batholiths and large intrusives of the Cordilleran region of the United States runs from gabbro to diorite, to quartz diorite, to granodiorite, to granite, to alkalic types. This is a time sequence

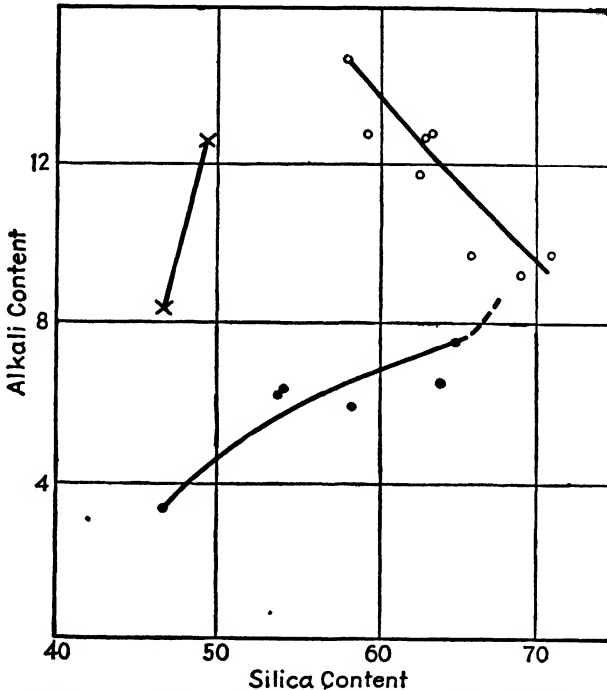


FIG. 111.—Diagram of the trend of differentiation in three rock series. The average rocks are different and the direction of variation is different in these three well-known masses. Spots, for Haystack Mountain, Montana. Circles for an alkalic mass in Essex County, Massachusetts. Crosses for the laccolith at Shonkin Sag, Montana.

as well as a petrographic series. A plot of the phases in 50 or more well-studied masses shows that there are a large number of other sequences that run from gabbro to alkalic types without including granite, and still others from granite to alkalic types without any indication of gabbro. Although the common sequence is from basic to silicic or alkalic rocks, a few series show alkalic rocks following the silicic rocks. The known series thus show a diversity in trend, which greatly complicates the problem of the origin of the series (Fig. 111). It is noteworthy, however, that probably more than half the masses studied had an average magma of generally basaltic nature as intruded, and the differentiates are olivinitic

¹ FENNER, C. N., The Katmai magmatic province, *Jour. Geology*, vol. 34, pp. 675-772, 1926.

at one extreme and quartzose and alkalic at the other. The other masses show much variety in the average magma, and in the differentiates formed—shonkinites, teschenites, syenites, etc. (see list of further readings).

The Bearing of Series on Classification.—Although students of analyses of igneous rocks find all gradations from one composition to another, this fact has not yet convinced all petrographers that classification must be wholly arbitrary, and that no natural divisions can be found. The study of series, like that of the curves of alkali and silica contents, shows clustering around two main stations. Figures 103 and 104 show the dominance of granite and basalt as clearly as the measurement of areas exposed (page 120). It may be hoped that eventually some standard will be set up for each of these main groups (and perhaps others), and a natural classification will be established, based on the nature and extent of a rock's deviation from these standards. This has not yet been done, however, and meanwhile the classification here given can be made to serve practical purposes better than some petrographers seem to think possible.

It should be added that different rock series vary for different reasons, and it is not always easy to determine what process has operated to produce any particular series (see page 252).

HISTORICAL GEOLOGY OF IGNEOUS ROCKS

Daly has compiled a table representative of the thousands of sequences of igneous rocks in many parts of the world. His summaries of the relations of rock types to geologic times are therefore probably indications of the earth's igneous history.

From the earliest pre-Cambrian to the present time basaltic lavas have been extruded most often and in largest volumes. On the other hand, in each geologic period the intrusive rocks of the basalt-gabbro clan have been subordinate to granite and granodiorite.

The pre-Cambrian was a period of especially intense igneous activity of all sorts. Granite batholiths developed in at least four distinct epochs in the pre-Cambrian and probably more. Plateau basalts were spread very widely about as many times.

Of the more restricted species the following are noteworthy: Anorthosites are pre-Devonian and mostly pre-Cambrian; alkalic rocks are largely post-Cambrian and in fact largely post-Paleozoic; the diaschistic dikes are mostly post-Cambrian. These restrictions may be due partly to erosion of those earlier rocks which are formed only in minor quantities and in border positions in orogenic belts.

The periodicity of the great batholithic invasions is emphasized by a tabulation of North American invasions (after Daly).

- | | |
|--------------------------|-------------|
| 1. Pre-Cambrian..... | Laurentian |
| 2. Pre-Cambrian..... | Algomian |
| 3. Pre-Cambrian..... | Keweenawan? |
| 4. Close of Devonian | |
| 5. Carboniferous-Permian | |
| 6. Jurassic | |
| 7. Early Tertiary | |
| 8. Miocene | |
| 9. Probably Pliocene? | |

Cycles of fusion of basalts have been outlined by Holmes¹ with even more striking though somewhat idealized rhythm. It must be admitted, however, that if the batholiths of all the world were determined as to age, a complete tabulation might show an invasion in progress somewhere in the world at almost any period or epoch. The apparent periodicity may be local.

ALTERATIONS OF IGNEOUS ROCKS²

Aside from the profound changes, which alter igneous rocks so far that they are classed as metamorphic, there are minor changes which may be very noteworthy. Some of these develop early and others long after the rock has cooled.

"Deuteric effects" are the synantetic minerals and the structures resulting from the reaction of early crystals and late magma residues. The sequence of reaction effects is suggested on page 242. Luxullianite and greisen, with associated cassiterite, are evidently a result of high-temperature alteration of granite. Some deep-seated kaolinization may be of similar nature.

Many crystalline rocks also have recrystallized with some freshening of the minerals and slight changes of texture, notably where they have been kept hot and acted on by hot waters, as they would be if intruded by other magmas. All glasses as old as the Archean seem to be devitrified.

Hydrothermal action proceeds by stages that are well recognized in connection with alteration near ore deposits especially near quartz veins—propylitization, sericitization and silicification. The change from basalt to "spilite" by the albitization of the basic feldspar is also attributed to hydrothermal action. Probably most amygdaloidal cavities are filled by hot solutions.

Weathering early produces clouds of kaolinite in the feldspars and may hydrate, or carbonate, or oxidize other minerals. It characteristically softens the rock and loosens the grains, increasing the volume unless much is leached out.

Dynamic effects, if not so great as to make the rock metamorphic, may be recognized in the bending of biotite grains and the bending or

¹ HOLMES, ARTHUR, Radioactivity and the earth's thermal history, Part 5, *Geol. Mag.*, vol. 62, p. 538, 1925.

² See Table XVI, page 306.

breaking of twinned plagioclase crystals. Mylonitization is a more complete crushing of an igneous rock to a fine granular mass, which, if it has not developed the usual banding or schistosity, may still be much like an igneous rock.

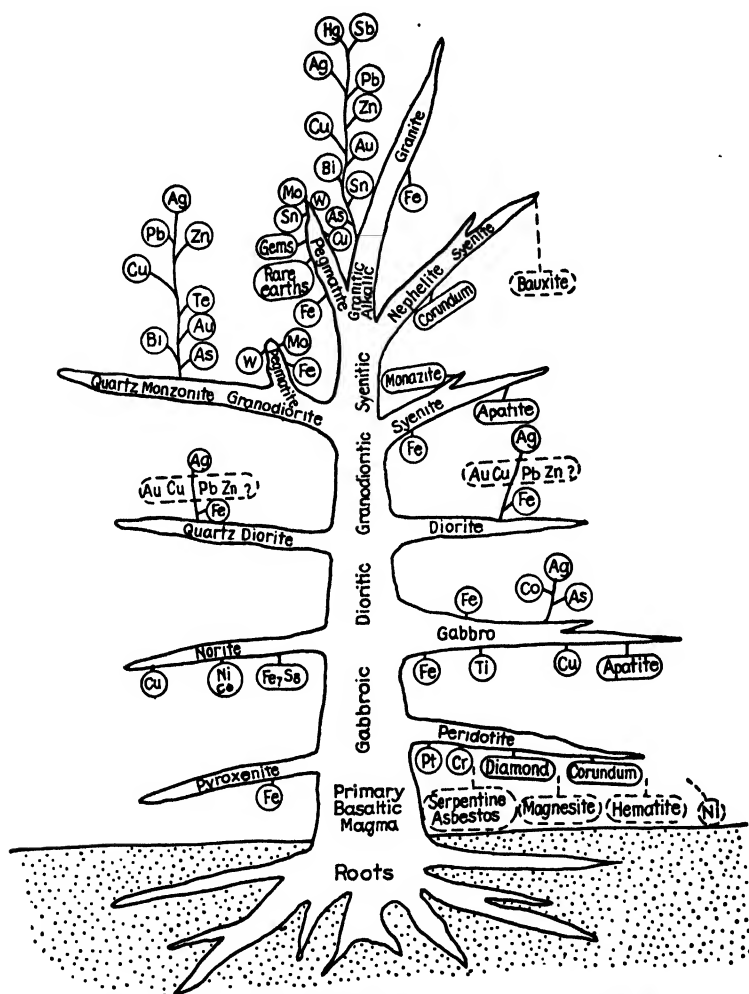


FIG. 112.—The relation of igneous rock clans to economic deposits. The elements that branch upward are found in veins, commonly in a sequence outward from the igneous rock; the elements hanging down like fruit are segregated; those falling to the ground are weathered, "rotten rock."

ECONOMIC MATERIALS RELATED TO IGNEOUS ROCKS

1. *Rocks used directly.* Nearly all species of igneous rocks may be used as structural material, and some are ornamental, the demand for

nem depending on the changing fashions. Granites are probably most used, but the darker kinds of stone are probably just as strong and desirable, except for their somber appearance. Granite is much used, also, as a monument stone. For concrete aggregate any igneous rock is excellent unless weathered badly. For macadam road metal, trap rock is best. Volcanic ash and pumice are excellent abrasives, and ash can be mixed with lime to make a "pozzuolan cement."

2. *The pegmatites* contribute a number of valuable minerals—mica, feldspar, and the gem stones, tourmaline, topaz and others.

3. *The ores* of most metals are related to igneous rocks, and the kind of rock very strongly indicates the metal to be found. The constituents of the earth may be classified as petrogenic and metallogenic according as they are found concentrated in rocks or only in the ore deposits, but the accidents of ore deposits emphasize the fact that the materials of the ores are largely derived from the rock magma. Figure 112 is an attempt to show the relation graphically (see also Fig. 141). The rocks from which ore may be formed by superficial alteration and enrichment are termed "protores."

Usually there are other characteristic associations of ores with particular rocks, such as the well-known occurrence of gold ores near the phonites at Cripple Creek, Colorado. Metallization is often related to late stages of magmatic action, pegmatitic or lamprophyric.

An important generalization also relates the type of ore deposit to the degree of erosion of a batholith. There is a concentration of mineralization near the apex of batholithic cupolas or the slightly truncated stocks. (See the list of further readings.)



FIG. 113.—Surface of a recent lava flow southwest of the crater of Kilauea. (*Photograph by H. S. Palmer.*) The surface is ropy, the pahoehoe type of lava. At times of volcanic eruptions, igneous rocks of this sort can be seen in the process of formation.

PART III

THE PETROLOGY OF IGNEOUS ROCKS

NATURE OF IGNEOUS MAGMA

Magma is a natural fluid in or on the earth, generally very hot, made up largely of a mutual solution of silicates, with some oxides and sulphides, etc.; it nearly always contains also water and other gases held in solution by pressure, but the heat is the main factor in its liquidity. So far as known, no magma is solid—not even a weak gelatinous solid.

Spurr¹ was led to believe that some magmas are gelatinous, and that all must retain water under pressure. This is not acceptable, though it is admitted that magma reaching the surface may be somewhat different from that at great depths.

Commonly the magma has been spoken of as "molten rock," but rock artificially melted under surface conditions differs from natural magma in a rather vital point—it contains no such quantity of water as is held in magma by great pressure in the earth. In a broad way magma may be a naturally molten rock, but even the lavas of volcanoes after eruption rapidly lose some of the characters of deep magmas.

We have no analyses of magma. The igneous rocks analyzed may approach the magma in composition but have lost at least a part of the volatile matter that is characteristic. Lacroix has emphasized this loss by a description of lherzolite, which has a metamorphic aureole of borated and alkalic minerals. The magma must have had boron and alkali, but the lherzolite has hardly a trace now.² Possibly some chilled pitchstone, which contains as much as 10 per cent of water, may give us our best notion as to the composition of magma in deep chambers.

Although the term magma primarily denotes a liquid, it may well include those cooling fluids in which partial crystallization has occurred, so long as the mass has not developed notable rigidity.

The history of a magma begins with its generation as a fluid and ends in its solidification by either crystallization or chilling to a glass. Magmatic history includes solution, transformation, emanation, movements, crystallization, reaction, differentiation, and other processes.

GENERATION OF MAGMA

Magma in general is a fluid. As the earth as a whole is rigid, we commonly think of the formation of a magma as a process of fusion.

¹ "Ore Magmas," pp. 140, 794, and 800, McGraw-Hill Book Company, Inc., 1909.
² Lacroix, *Le granit des Pyrénées*, Carte geol. de France, vol. 10, No. 64, 1898-90, and vol. 11, No. 71, 1899-1900.

Certain theories of earth development relegate that fusion to such an early stage that the magma is residual rather than generated in the ordinary sense. A calculation of the necessary cooling without crystallization of viscous magma to make it as rigid as the earth is known to be shows very little likelihood that any large molten layer in the earth is residual fluid below a few hundred kilometers (see pages 182 to 184).

Magma may be generated by three distinguishable processes or combinations of them: (1) increase of heat, (2) decrease of pressure, (3) change of composition, the only probable change being an addition of water.

Although it is known that the addition of water will reduce the temperature of fusion (page 154), it is questionable whether water or any other constituent that will effectively change the composition can penetrate a rock mass below the zone of fracture (page 212). The streaming of gases in a magma (page 218) may so increase the activity of solutions as to cause "gas fluxing" and assimilation, and these processes in one sense produce more magma, but the processes of assimilation are here distinguished (pages 225 to 230).

The relation of the temperature of fusion to the pressure is discussed on pages 152 to 156. Although the data are not absolutely conclusive, it seems reasonably certain that pressure raises the melting temperature of rocks and that release of pressure on some hot rocks deep in the earth will cause fusion. The geologic causes of release of pressure are, first, the erosion of the land and, second, the arching of some competent rock structures during orogenic movements. Either of these might produce a local mass of magma, but it is difficult to cite cases in which evidence is strong for this result. The release of pressure by erosion is slow, and the descending geothermal gradient might cool the rocks below almost as fast as the pressure was released. The erosion needed to fuse a basalt is also found to be surprisingly great—of the order of 25 kilometers of rock.¹ Again, the orogenic arching must be of comparatively short duration, and the generated magma would resolidify unless in that short period it moved to a higher position in the lithosphere. Vogt estimates that a release of pressure in arching rocks a kilometer thick would change the melting point only 1° to 5°. Arching seems hardly probable as a general source of magma.²

The problem of magma generation thus becomes largely a problem of heat supply.

MAGMATIC HEAT

No mineral or rock remains solid above its fusion temperature under the prevailing pressure. Estimates of the fusion points of minerals and

¹ BOWEN, N. L., "Evolution of Igneous Rocks," p. 315, Princeton Univ. Press, 1928.

² VOGT, J. H. L., Magmatic differentiation of igneous rocks, Jour. Geology, vol. 30, p. 614, 1922.

rocks made by different laboratories show some disagreement (see pages 152 to 154). Some minerals, it should be noted, are transformed before melting.

The temperatures actually observed in a number of liquid lavas are around 800° to 1200°C. Lava with crystals growing in it has been observed at 1000°C., and at Kilauea lava still flows slowly at 600°C. All these observations were made on andesitic or basaltic lavas. For rhyolites and for deep seated magmas the evidence is less direct, but the temperatures are probably lower. There seems to be little evidence that any magma in the earth has been hotter than the surface lavas, perhaps none so hot, for the lavas observed in craters may be somewhat hotter than is necessary to keep them liquid, because of local heating where gaseous emanations are burning.

Judging from the temperatures of rock melts, from the known changes in the forms of silica, from the contact effects produced by the magmas, and from experimental work with hydrous melts, it seems likely that magmas come to place at 870° to 1250°C.,¹ and start to crystallize at depth around 800° to 900°C. The hydrous siliceous magmas remain fluid to rather low temperatures, but only a very small percentage is left to crystallize below 573°C.—the transformation point from alpha to beta quartz. Late basic dike magmas may also remain liquid to low temperatures.²

Sources of Heat.—The heat found in magmas and lavas is differently interpreted according to different theories as to the origin of the earth. Some heat may be left from an originally intensely hot gaseous globe, now greatly cooled. Various calculations and observations seem to cast doubt upon this idea of the original heat.

Under the planetesimal hypothesis the heat may come from several sources.

1. The nucleus and planetesimals may have been warm.
2. Their accumulation or infall upon a nucleus developed heat of impact.
3. Heat results from compression of the core by the pressure of later planetesimals accumulated around it.
4. Heat may result from reaction between different planetesimals.
5. Radioactive elements generate heat as they break down to smaller atoms.
6. Heat is developed by friction along faults and even by rock flowage and viscous magma flowage.

¹ LARSEN, H. S., The temperatures of magmas, *Am. Mineralogist*, vol. 14, pp. 81-94, 1929. Higher estimates were made by Vogt in *Jour. Geology*, vol. 31, pp. 407-419, 1923.

² SPURR, J. R., Basic dike injections in magmatic vein sequences, *Geol. Soc. America Bull.*, vol. 36, pp. 553-564, 1925.

7. "Porous plug expansion"—the escape of magma from a chamber through an orifice—generates heat.¹

8. Heat reaches the earth from the sun.

See further readings listed.

Some losses of heat may tend to counterbalance these sources. Some reactions absorb heat, and the earth may radiate as much heat as it receives from the sun. Nevertheless several of these sources may furnish heat which will, without itself generating magma, locally change the geothermal gradient and possibly facilitate heating at depth.

The student should carefully distinguish between *pressure* and *compression*. The compression of air in a pump develops heat, but if the air is pumped into a container and kept under pressure, that continuous pressure does not raise the temperature.

The rate at which heat was developed by the infall of planetesimals is not settled. If planetesimals were large and struck so frequently and with such velocity that the heat of impact accumulated faster than it was radiated into space, the surface of the growing earth might have been molten. The originators of the planetesimal hypothesis did not believe the growth was rapid enough to have this effect (see page 182).

The heat of compression as planetesimals accumulated has been estimated as reaching 20000°C. in the interior of the earth. Temperatures near that estimate are so far beyond any that we can reach in the laboratory that we have no idea as to their effects on the rocks.

A gram of radium emits each hour from 100 to 135 gram calories, according to various estimates. Other radioactive elements even potassium also give heat. The heat emitted by the several elements differs widely; a calculation for average rock gives:

Element	Heat Generated in Calories per Gram of Rock per	
	Year	
Uranium.....	4.74	$\times 10^{-6}$
Thorium.....	3.45	$\times 10^{-6}$
Potassium.....	3.22	$\times 10^{-6}$

No laboratory temperatures or pressures affect this heat generation in any way, so that it seems likely that any such elements deep in the earth generate similar quantities of heat. Joly questions if uranium was always constant. An isotope may have generated more heat in early times.²

It is known, however, that pressure may change or reverse some reactions, and it is not impossible that atoms of uranium and radium may be *built up* under such pressures as occur deep in the planets or stars with absorption of heat (see page 159). To emphasize our uncertainty it must be admitted that for all we *know* the core of the earth may be at

¹ ADAMS, L. H., A physical source of heat in springs, Jour. Geology, vol. 32, pp. 191-192, 1924.

² Recent work by Mühlhoff indicates that the heat evolved by potassium is much less than here estimated.

absolute zero (or gaseous at 200000°C. Few think either temperature is likely.

Radioactive elements are known to be more abundant in silicic than in basic rocks.¹

RADIUM AND THORIUM, GRAMS PER GRAM OF ROCK
(Modified from Poole and Joly)

Silicic rocks.....	Ra = $3.0 \pm 4 \times 10^{-12}$	Th = 2.31×10^{-5}
Intermediate.....	Ra = $2.2 \pm 1 \times 10^{-12}$	Th = 0.61×10^{-5}
Basalts.....	Ra = $1.3 \pm 1 \times 10^{-12}$	Th = 0.60×10^{-5}
Ultrabasic rocks...	Ra = 0.8×10^{-12}	Th = 0.61×10^{-5}

Further, we have good indications that granite is most abundant near the surface and that deeper rocks grade into basalts and probably still deeper into peridotites. Such rocks would have about one-third or one-fourth as much heat generated in them by radioactivity as the granites. Probably at greater depths this source of heat is almost negligible. It is estimated that a layer of 20 kilometers of such rocks as occur at the surface would be equivalent to all the radioactive rocks of the earth.

In summary of radioactive heat supplies it should be noted that when discovered they seemed to be amply sufficient to account for all the heat we find in the earth's crust. The supply from uranium and thorium alone was disturbing to some of the older theories, and as the heating power of potassium adds considerably to the total, the idea of a cooling earth is being completely abandoned by some prominent petrologists. (See the further readings listed.)

The heat of friction must be very great in those earth movements that thrust great bodies of rock over other rocks for several miles. In depth, where the temperature is already high, the added heat of such friction may produce a magma along the plane of thrusting. Geographically there is a noteworthy correlation of magmatic invasion and thrusting in areas of active mountain building, though few magma bodies occupy the thrust planes. If once produced along a thrust plane, however, a magma would by its very nature tend to obliterate the evidence of its origin, and no very clear cases are known in which a particular magma is definitely ascribed to friction. A few flinty, crushed rocks grade into black glassy-looking rocks that are microscopically isotropic and have flowed, though possibly not actually fused (page 413). These are very suggestive.

The local development of heat, which might produce a local reservoir of magma as distinct from a regional or universal magma, may be due to (1) a local abundance of radium, or (2) a local friction from earth move-

¹ POOLE and JOLY, Radioactivity of basalts and other rocks, *Philos. Mag.*, vol. 48, pp. 819-823, 1924. HOLMES, A., Radioactivity and the earth's thermal history, *Geol. Mag.*, vol. 62, p. 68, 1915; vol. 62, p. 509, 1925. PIGGOT, C. S., Radium in rocks, *I. Am. Jour. Sci.*, 5th ser., vol. 17, pp. 13-34, 1929.

ments, or (3) a local rise of heat in geosynclines. Holmes¹ suggests a geosyncline filled with sediment as the explanation of a local concentration of a great thickness of rock with more radioactivity than average basalt.

Maintenance of Heat.—Small masses of intruded magma ordinarily cool by conduction of heat into the wall rocks and solidify in a comparatively short time. The feeders below volcanic craters, however, seem to remain molten and hot for hundreds if not thousands of years. Some process must be supposed to supply heat to them more or less continuously. The heat may be in part a result of burning and reaction of gases in the crater lake, but more likely most of it is supplied from some larger mass of magma in a hot deep-seated reservoir. The mechanism of its supply to the volcano is probably either convection or a "streaming" of gases.

Convection in a volcanic throat may be due to a change in density of the lava on cooling, but fully as important changes in density probably arise from local vesiculation or the local growth of crystals in the magma. This is a two-phase convection (see pages 206 and 251).

Gases dissolved in magmas have a tendency to "stream" toward the points of least pressure (page 218). If, as should be expected, they rise from deep, highly heated regions, they no doubt carry a certain amount of heat.

DATA AND THEORY AS TO THE INSIDE OF THE EARTH AND ITS RELATION TO MAGMAS

TABLE VIII.—DATA MUCH USED IN PETROLOGIC CALCULATION

1 mile	= 1.609 kilometers	1 kilometer	= 0.6214 mile
1 inch	= 2.54 centimeters	1 centimeter	= 0.3937 inch
1 pound	= 0.4536 kilogram	1 kilogram	= 2.205 pounds
1 cubic foot of water weighs	62.5 pounds.		
1 gallon of water weighs	10 pounds.		
Radius of the earth is	3,980 miles = 6371 kilometers.		
1 atmosphere of pressure	= 15 pounds per square inch.		
1 megabar	= 1,000,000 dynes per square centimeter		
	= 1.0197 kilograms per square centimeter		
	= 0.9869 atmosphere		
	= 14.8 pounds per square inch		
	= 75.0068 centimeters of mercury		
Critical temperature of water	= 374°C.		
Atomic weights (see Am. Chem. Soc. Jour., May, 1931).			
Melting points of minerals (page 152).			
Conductivity of rocks (page 151).			
Thickness of a good thin section	= 0.03 mm.		
Diameter of the earth	= 7,918 miles = 12,743 km.		
Volume of the earth	= $259,886 \times 10^6$ cubic miles		
Volume of a 10-mile crust	= $1,633 \times 10^6$ cubic miles		
Volume of ocean	= 302×10^6 cubic miles		
Mass of the earth	= $6 \pm \times 10^{21}$ metric tons		
Mass of ocean	= 1.27×10^{18} metric tons		
Mass of atmosphere	= 7.3×10^{15} metric tons		

¹ HOLMES, ARTHUR, Radioactivity and the earth's thermal history, *Geol. Mag.*, vol. 53, pp. 271-272, 1916.

Area of ocean	= 144.7×10^6 square miles
Area of lands	= $55. \times 10^6$ square miles
Area of earth	= 199.7×10^6 square miles
Mean elevation of continents	= 2,411 feet above sea level

APPROXIMATE PRESSURES OF OVERLYING ROCK AT SEVERAL DEPTHS IN EARTH¹

Kilometers below surface	Megabars
0.....	1
1.....	270
5.....	1,350
10.....	2,700
20.....	5,500
50.....	15,000
100.....	32,000
1,000.....	375,000
2,400.....	1,000,000
6,370 Center of the earth	3,180,000

¹ After ADAMS and WILLIAMSON, Smithsonian Rept. for 1923, p. 259, 1925.

SOURCES OF MAGMA

Magmas come from "below." The justification of the opening statement of this paragraph may not be complete until the discussion has run on for nearly a hundred pages; see especially pages 185 to 190. The term "below" covers a vast amount of ignorance and speculation. Nevertheless, although the discussion of sources is theoretical in a high degree, the choice of theories may affect the petrographer's observations of fact so far that it is worthy of considerable attention. An attempt will be made to separate rather sharply the records of observation from the theories based on them, so that the student will not be misled into thinking that the prevailing theory is universally accepted.

The facts are presented in somewhat abrupt and disconnected form as a basis for the theories and arguments to follow. From the data more or less accurate are built a series of hypotheses from which those worthy of retention as *working* hypotheses may be selected.

GEOTHERMAL GRADIENT

Data.—The geothermal gradient has been observed at various places to depths of several thousand feet. Borings have reached 10,500 feet. The general average increase of temperature with depth is about 32°C. per kilometer. Certain corrections may raise this to 35°C. per kilometer, and there is reason to believe that near the bottom of deep wells the rise is faster than near the earth's surface. For most purposes the gradient may be estimated as a little over 1°C. per 100 feet.

Theory.—Beyond the short range of observation the gradient is variously estimated. If the observed gradient were extrapolated as a

straight line, temperatures at about 100 km. would be 3000°C , and at the center of the earth more than 200000°C . With such temperatures common shales at depths of 10 or 15 miles should be fused, but structural studies show that many shales have been buried at such depths without any sign of melting. Lavas reaching the surface in some places come in such great volume and so fast as to indicate a rather deep source, but they have not such great heats as the estimate above—more nearly 1000° to 1300°C . Nearly all petrologists therefore believe that the gradient changes, and hardly reaches above 1300° at depths of 100 to 200 km. (Fig. 114).

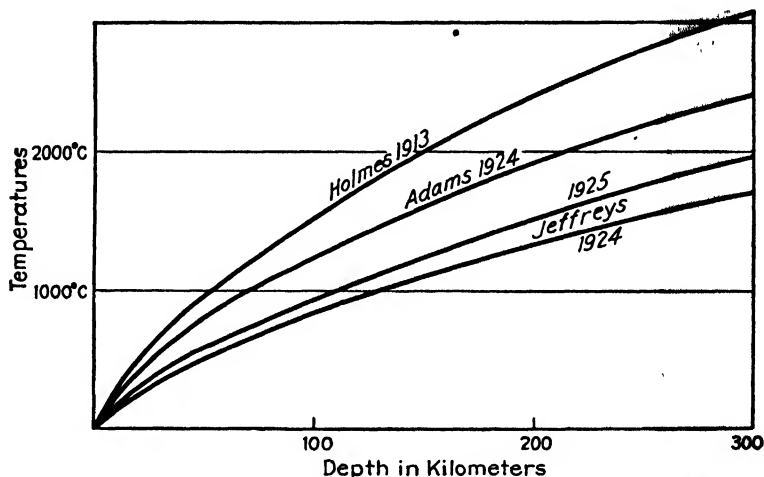


FIG. 114.—Recent estimates of the temperatures of the earth at moderate depths.

If the heat in the deeper earth is largely due to compression of rocks like those in the crust, and has not been much modified by other factors, the temperature of the center can be calculated as about 20000°C . If the core of the earth is denser than the crust, partly because of a concentration of heavy metal, the temperature calculated should be correspondingly reduced. The factors affecting estimates are many and uncertain. It is unknown whether the earth ever was molten or had a molten external layer; whether metal segregated at the center; whether the crust foundered progressively. The heat supplies are probably largely those of compression and radioactivity, but there may be large modifying factors quite unknown; some reactions may absorb some of the heat (page 159).

The heat conductivity of rocks is relatively low.

The observed gradient is not determined by conduction of central heat alone; this source is modified by igneous invasion near the surface and by the distribution of radioactive materials. It is to be expected that the gradient will vary erratically in time and place, on account of

local and periodic intrusion of magma, and the differences in conductivities of different rocks. This agrees so well with the differences in the observed geothermal gradient that any curve sketched should be understood as only a broad average, much smoother than the actual temperatures warrant.

APPROXIMATE CONDUCTIVITIES

Silver.....	1 000
Lead.....	0 0836
Quartz.....	0 0158
Marble.....	0 008
Granite.....	0 006
Sandstone.....	0 005
Basalt.....	0 004
Syenite.....	0 004
Water.....	0 0013
Cork.....	0.00013

Altogether it is doubtful if any sound conclusions as to the former temperatures of the earth or its present temperature below four or five miles can be based on the *observed* gradient. The study must be based largely on more remote data.

Adams¹ recently calculated the probable curve of temperatures to a depth of 300 kilometers on the several assumptions that (1) the surface of the earth, if not all of it, was once hot enough to be molten; (2) the surface cooled and the deeper temperatures were for a time controlled by convection, which was ended by increase of viscosity or by crystallization and crust foundering; (3) when convection ceased and cooling began by conduction, the curve of temperature was nearly that of the fusion curve to great depths; (4) the age of the oldest rocks (from lead-uranium ratios) is about 1,600 million years; (5) the upper shell of earth is granitic and some deeper shells are basaltic and peridotitic; (6) the heating effect of atomic disintegration is allowed for. He found that various assumptions as to early surface temperatures had little effect on the calculation of present temperatures, these being dependent on the deeper primitive temperatures. His estimate is shown in Fig. 114 as a first approximation.

Adams's calculations seem to ignore one of the effects of igneous intrusion, namely, that magmas probably brought heat supplies from depths of perhaps 100 km. or more to or near to the surface—from regions where temperatures of 1300°C. are normal to places where temperatures less than 500°C. are normal. The observed geothermal gradient is much steeper than it would be without this transfer of hot magma. It seems evident then that the curve of cooling without such transfer should show lower temperatures near the surface than the actual observed gradient. Probably the whole curve as shown by Adams is at too high a temperature.

The fusion-pressure curve also needs study (see pages 153 to 156). Pressure holds water and other gases in solution in magmas at high temperature, thus keeping the fusion point much lower than it would be if the gases escaped. Probably Adams's estimates of the fusion temperatures are somewhat too high for this reason. Furthermore, Adams shows the fusion curve of peridotite as a straight line, although there seems to be plenty of evidence that nearly all substances show a curve with small temperature rise at high pressure. Finally, the fusion curve plotted by Adams would not come very close to the geothermal curve until carried to nearly 500-km. depth;

¹ Adams, L. H., Temperatures at moderate depths within the earth, Washington Acad. Sci. Jour., vol. 14, p. 468, 1924.

but the yielding nature of the crust at depths a little more than 100 km., shown by the data of geodesy, suggests that the curves approach each other at depths of perhaps 60 km.

The curve shown by Adams reaches 2000°C. at a depth of 200 km.; but few observed lavas have so high a temperature, and most students, as stated above, seem to think the rocks at such depths are little if any above 1300°C.

FUSIBILITIES AND FUSION-PRESSURE CURVES

Data.—The data of the table are compiled from various sources, which do not entirely agree. The Geophysical Laboratory of the Carnegie Institution has done some especially careful work, and its data are selected in most cases.

TABLE IX.—MELTING POINTS OF ROCK MINERALS

Mineral	Composition	Temperature degrees Centigrade
Periclase.	MgO	2800
Spinel.	MgAl ₂ O ₄	2135
Corundum.	Al ₂ O ₃	2050
Zircon.	ZrSiO ₄	1900
Forsterite.	Mg ₂ SiO ₄	1890
Leucite.	KAl(SiO ₃) ₂	1820–1830
Cristobalite.	SiO ₂	1713 ± 10
Tridymite.	SiO ₂	1670 ± 10
Rutile.	TiO ₂	1700
Fluorapatite.	Ca ₄ (CaF)P ₂ O ₁₂	1650
Clinoenstatite.	MgSiO ₃	1557 incongruent
Anorthite.	CaAl ₂ Si ₂ O ₈	1550 ±
Nephelite (Carnegieite).	NaAlSi ₃ O ₈	1526
Diopside.	CaMg(SiO ₃) ₂	1391
Calcite (pressure above 1,000 atm.).	CaCO ₃	1340
Titanite.	CaTiSiO ₅	1220
Magnetite.	FeO.Fe ₂ O ₃	1185
Micas.	Variable	1130–1230
Orthoclase.	KAlSi ₃ O ₈	1170 incongruent
Pyrrhotite.	Fe ₇ S ₈	1185 ± 15
Albite.	NaAlSi ₃ O ₈	1100 ± 10
Fayalite.	Fe ₂ SiO ₄	1065
Acmite.	NaFe(SiO ₃) ₂	990 incongruent
Other pyroxenes.	Variable	925–1400
Amphiboles.	Variable	1025–1085
Halite.	NaCl	801

It is well understood that many mixtures of minerals fuse at lower temperatures than single minerals (see page 234). Even more important is the fact that water vapor under moderate pressure is dissolved by molten minerals (and magmas) in quantities that greatly lower the melting temperature. It is estimated that each per cent of water added to a magma, up to several per cent, will reduce the melting temperature

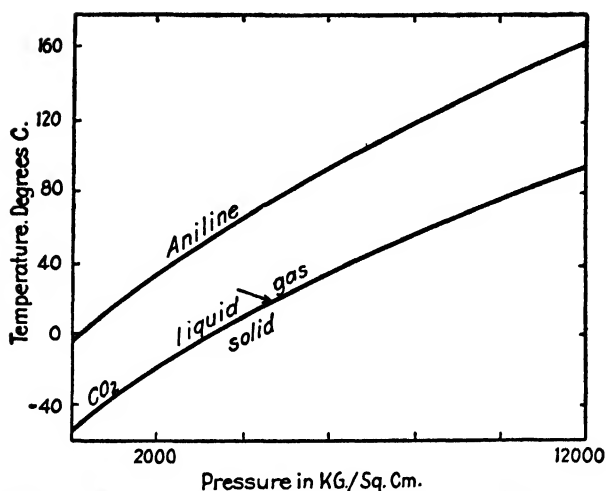


FIG. 115.—Fusion-pressure curves for aniline and for carbon dioxide. (After John Johnston.)

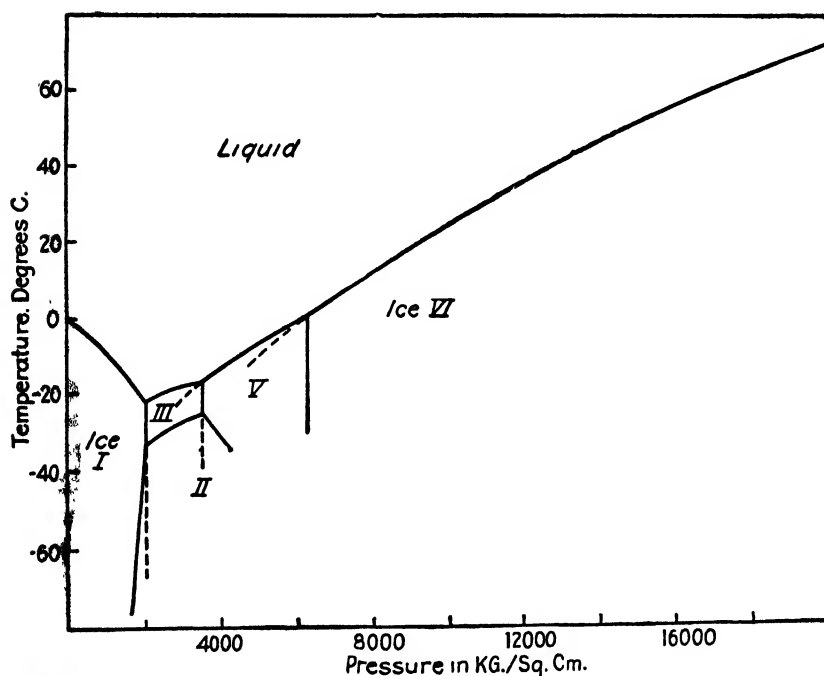


FIG. 116.—Fusion-pressure relations for liquid water and ice in five solid forms. (After Bridgman, *Jour. Geology*, vol. 23, p. 739, 1915.)

about 50°C. Morey says¹ that 0.1 per cent of water reduces the fusion point of anorthite 5°C.

The melting points, or better the melting ranges, of rocks in the laboratory at atmospheric pressure have thus relatively little significance, but as there has been some disagreement on the relative melting temperatures of different rocks it should be noted that the melting temperatures for basalts are about 200°C. higher than those for rhyolite.²

P. W. Bridgman, at Harvard, has tested the relation of pressure to fusion temperature for a number of substances. Unfortunately the temperature of fusion of rocks is so high that direct high-pressure experi-

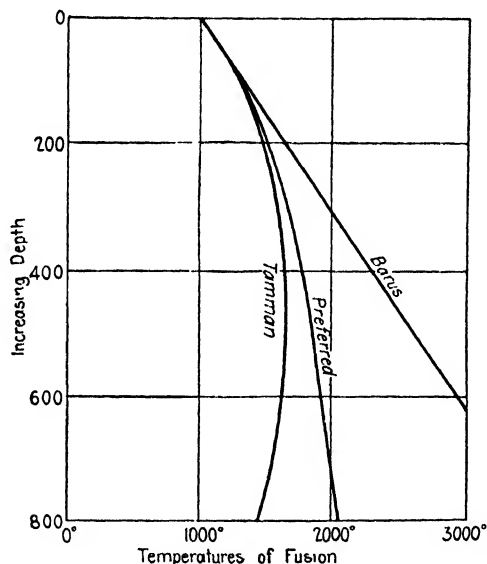


FIG. 117.—Diagrammatic contrast in the several extrapolations of the fusion-depth curve.

ments on such materials are very difficult. We must infer the behavior of rock melts from the behavior of others. Figure 115 for aniline and for CO₂ are typical of most melting curves. The melting points of nearly all substances are raised by pressure. The curve for ice is exceptional at low pressures, and no other substance except bismuth is known whose melting temperature falls as pressure is increased. Even ice is not an exception at high pressure;³ at 20,000 kg. per square centimeter ice would form from water hot enough to scald your hand! (Fig. 116). Note that each succeeding increment of pressure gives a smaller effect. The curves of Fig. 115 are convex upward.

¹ MOREY, G. W., The development of pressure in magmas as a result of crystallization, Washington Acad. Sci. Jour., vol. 12, p. 223, 1922.

² GREIG, SHEPARD, and MERWIN, Melting granite and basalt in the laboratory, Geol. Soc. America Bull., vol. 40, pp. 94-95, 1929.

³ JOHNSTON, JOHN, Some aspects of recent high-pressure investigation, Franklin Inst. Jour., pp. 7-10, January, 1917.

One other observation may be significant at higher temperatures and pressures. It has long been known that melts of basaltic composition contract on solidification.¹ This suggests that magmas will be solidified by high pressures, even as the aniline and CO₂ are solidified, but further data, especially on rocks, are much needed. Different investigators estimate the increase in temperature of fusion at values from 2.5° to 5°C. for an increase in pressure equivalent to a kilometer of rock.

Theory.—As investigators differ in regard to the effects of pressure on the fusion of rocks, as stated above, and as experimental pressures are equivalent to about 64 km. of rock whereas the center of the earth is

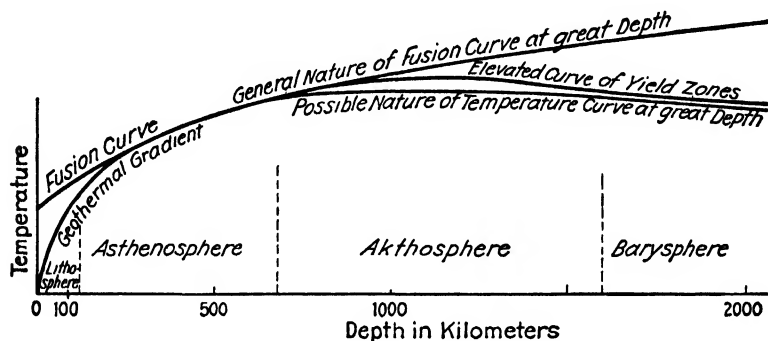


FIG. 118.—Possible relation of the fusion curve for rocks and the changing gradient at depths in the earth. (After Barrell, *Am. Jour. Sci.*, 4th ser., vol. 10, p. 508, 1925.) This suggests also the depths at which the strength of rocks might be reduced by high temperatures. Note that if the temperature curve was slightly raised it would cross the fusion curve, indicating a magma zone or substratum.

down about 6,400 km., the estimates are very uncertain, and the curve is extrapolated 100 times farther than it is determined by poor data (Fig. 117).

Barus early suggested that the relation of fusion and pressure could be extrapolated as a straight line to great depths.

Tamman has argued from altogether insufficient evidence that there is a maximum on the curve. Joly has made a similar suggestion, as if all rocks could be melted if the pressure was high enough. This suggestion is based on experiments with a few organic substances and is not a sound basis on which to oppose the data obtained on inorganic substances.

Recent work shows quite definitely that even gases could probably all be solidified at high pressures. Bridgman's accurate work on inorganic

¹ BARUS, CARL, *Igneous fusion and ebullition*, U. S. Geol. Survey Bull. 103, p. 25. 1893. Recent detail by Day, Sosman, and Hostetter, *Am. Jour. Sci.*, 4th ser., vol. 37, pp. 1-39, 1914, shows a contraction of about 10.9 per cent. The doubt cast on these experiments by Harker's observation of a glass border of a dike more dense than the crystalline equivalent ("*Natural History of Igneous Rocks*," pp. 158-163) cannot carry much weight.

material shows that the increase of temperature is slower at high pressures but shows no sign of cessation. We must assume that rocks decrease in volume when they solidify as most other substances do.

Possibly one of the best checks on the pressure-fusion curve is a comparison of that curve with the geothermal gradient (Fig. 118).

It is noteworthy that the present temperatures may be divided into several components (Fig. 119). Near the surface the heat might be the sum of the heat conducted out from the hot core, the heat of magma intrusion and the heat of atomic disintegration, minus the heat radiated. It is said that three-fourths of the heat loss is supplied by atomic disintegration.

Summary of Relation of Magmas to Heat and Pressure.—The geothermal gradient as best estimated gives temperatures as high as the surface fusion temperature of basalt, well within 100 miles below the surface. Pressure may keep the fusion curve as high as the temperature

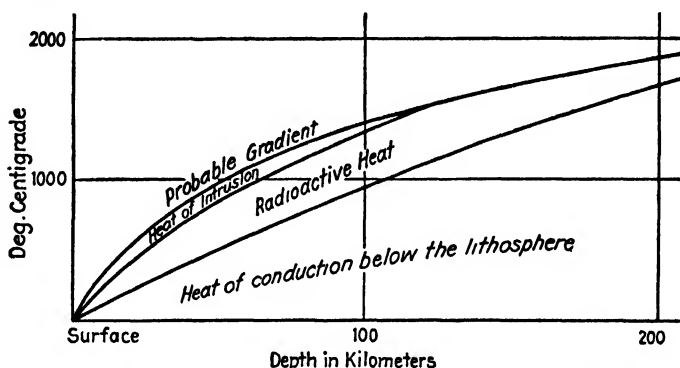


FIG. 119.—Rough diagram to emphasize the fact that the temperatures found near the surface are not all a result of conduction from the greater depths.

curve, but at depths of 50 to 100 km. the two curves approach very closely, and from there to great depths it is likely that the balance is very delicate. Any heat added will be likely to be the latent heat of fusion, and magma will form. The added heat may come from atomic disintegration, dynamic action or friction. Any release of pressure could similarly produce magmas in this zone of the earth. A question may arise as to the effectiveness of some of the processes noted here. Any particular one might be found inadequate to explain some particular case. For example, volcanoes are known in several regions where dynamic action could hardly explain the production of magma. In some regions one factor may be prominent, in other regions another. No uniformity in all parts of the earth need be assumed. (See also the further readings listed.)

DENSITIES, PRESSURES, AND COMPRESSIBILITIES

Stresses Produced by the Weight of Rocks

Data.—Physicists calculate that the density of the earth is about 5.52 and some calculations on the effect of mountains yield similar values. In contrast to this, the average density of the crust is about 2.79, and the densities of rocks of continents and ocean basins are about 2.75 and 2.87, respectively, if the samples taken can be considered representative. This means that the density at the center must be larger

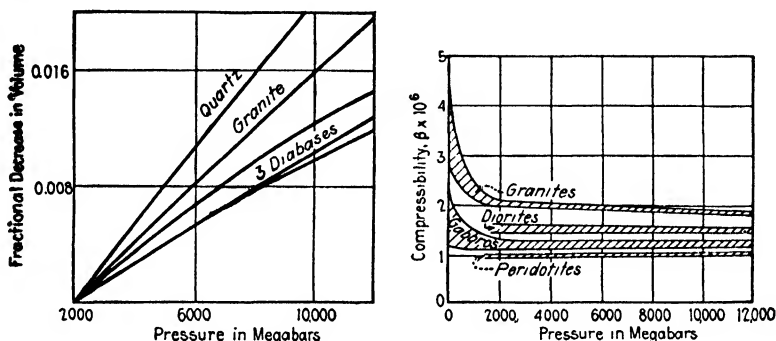


FIG. 120.—The compressibilities of common rocks. (After L. H. Adams and others.) Note that the lines of the diagram at the left are curved, not straight, lines. The width of the shaded areas indicates the variation to be expected within a class of rocks. At low pressure the variation is dominated by variation in the compactness of rocks, but at high pressures porosity has very little effect on compressibility.

than 5.52, and as the evidence is strong that surface rocks extend inward for a considerable distance, the density of the center may be estimated at 8 or 10—possibly even more.

Certain other observations confirm the general idea of a heavy center in the earth. Most of the deeper rocks in the crust are igneous rocks and occur in large bodies, and the large bodies of igneous rocks that are known from top to bottom are *gravitatively differentiated*. It is therefore quite certain that under the deep igneous rocks that we know, there are heavier segregations.

The compressibility of a few substances has been tested up to 20,000 megabars. Any mineral put under great pressure can be forced into less volume and thus becomes more dense.¹ Such compression has been referred to above as a source of heat in the earth. How much can the volume be changed? The compressibility of practically all substances decreases with increasing pressure.² The laboratory data apply only to rocks at moderate depths—around 50 kilometers (Fig. 120).

¹ A study of metamorphism has shown that minerals recrystallize at depth to other minerals, usually with greater density. This is not the change here considered.

² ADAMS and GIBSON, The elastic properties of certain basic rocks and their minerals, Nat. Acad. Sci. Proc., vol. 15, pp. 713-724. 1929.

TABLE X.—COMPRESSIBILITIES OF MINERALS AND ROCKS

(After L. H. Adams and others)

	Compressibility ($\beta \times 10^6$) per megabar		
	At low pressure	At 2,000 megabars	At 10,000 megabars
Minerals			
Quartz	2 70	2.61	2.27
Orthoclase	1 92	1.88	1.68
Labradorite	1 48	1 44	1 28
Augite	1 07	1 05	0 98
Calcite	1 39	1 39	1.39
Olivine	0 85	0.84	0 79
Garnet	0 61	0 60	0 57
Magnetite	0 55	0 54	0 51
Rocks			
Granite	2 12	1.88
Granodiorite	1 83	1.66
Basalt	1.88	1 55
Gabbro	1 20	1 17
Pyroxenite	1 03	1 03
Peridotite	0.97	0.77
Marble	1 41	1.41

The compressibility involved in the transmission of earthquake waves, "seismically effective," may be considerably less than the compressibility by the high-pressure method.¹

Theory. A. Compression and the Material inside the Earth.—If common minerals could be progressively compressed by additional pressures, at the rates shown by the first 20,000 megabars, up to 3,000,000 megabars, which is the approximate pressure at the earth's center, they might attain densities of 9.00. It will be noted, however, that this estimate carries the curve one hundred fifty times as far as the actual experiments have gone. It is very doubtful if the extrapolation is justified. The important point to notice in the experimental data is not the absolute amount of compression at certain pressures, but the fact that the compressibility is less at high pressures for most of the materials tested. This must be true theoretically also; for even when the pressure is almost infinite, we cannot think that the volume could be reduced to zero; something must remain and occupy space.

The final conclusion is that compression probably could not make rocks like those at the earth's surface so dense as 9.00 at the pressure prevailing in the earth's core. The known high density is due to a

¹ DALY, R. A., Nature of certain discontinuities in the earth, *Seismol. Soc. America Bull.*, vol. 20, p. 46, 1930.

difference of material. If, as suggested by some other data, the material of the earth is comparable to that of meteorites, it seems likely that the metallic nickel-iron has been largely segregated toward the deeper parts of the earth, leaving very little in the known igneous rocks.

B. Is There Contraction or Condensation at Depth?—The apparent shortening of the earth's crust that accompanies diastrophism is enormous. Barrell¹ estimated the shortening at one per cent of the earth's circumference in Cenozoic time, several per cent since the Cambrian, but there is much disagreement on these estimates. It seems that the inner part of the earth is changing its volume² by some process which is continuous, and which since early pre-Cambrian time, shows little tendency to fade out into any stability. Two such processes are suggested; one a cooling, which is not easy to reconcile with the data of radioactivity and the fact that cooling should after a time become slower; the other a molecular or atomic condensation, something akin to the slow progressive transformation seen in radioactivity, but of a reverse nature, building more complex atoms and possibly absorbing heat.³ There need be no doubt that the evolution of the chemical elements is a reversible process, though physicists are not yet agreed that the pressures such as those in the earth would reverse any atomic disintegration.

The geologic record in certain places shows a periodicity of diastrophism, suggesting the gradual accumulation of forces until they reach an intensity that results in a yielding of the resisting material. Under the idea of a shrinking core there might be postulated a strong shell of rock between the core and outer crust. This shell would accumulate stresses and yield only a little elastically until the stresses became too great, then yield more fundamentally, producing a diastrophic revolution. The whole idea of periodicity, however, is questioned and may be based on observations in too small an area.⁴ World-wide studies of diastrophism show it to have been widely distributed through geologic time. Probably no period can be said to be free from diastrophism all around the earth. Contraction of the core of the earth is still a good working hypothesis, but it is neither proved nor disproved.

Under the idea of a periodic fusion Joly and Holmes find an explanation of the periodicity in the slow accumulation of radioactive heat, followed by continental drift and oceanic cooling (pages 182 to 185).

¹ BARRELL, JOSEPH, Geologic relation of earth condensation, *Am. Jour. Sci.*, 5th ser., vol. 10, p. 414, 1925. See also, C. K. LEITH, "Structural Geology," new ed.; JEFFREYS, "The Earth"; and T. C. CHAMBERLIN, *Jour. Geology*, vol. 32, p. 716.

² The drift of continents might explain thrusts without condensation (see p. 173).

³ BARRELL, JOSEPH, The spheres of the earth and their effects on the lithosphere, *Am. Jour. Sci.*, 5th ser., vol. 10, p. 502, 1925.

⁴ BERRY, E. W., On correlation, *Geol. Soc. America Bull.*, vol. 36, pp. 266-267, 1925. F. P. SHEPARD, To question the theory of periodic diastrophism, *Jour. Geology*, vol. 31, p. 599, 1923.

C. The Pressures on Rocks at Various Depths in the Earth.—This pressure can be roughly calculated from the densities, though it is necessary to make some assumption as to the distribution of materials in the earth (page 178). The different assumptions make no difference that is vital to the general discussion. The curve of pressures rises from 1 megabar at the surface to about 3,180,000 megabars at the center and may be estimated at various depths (see page 149).

D. The Stresses in the Interior of the Earth.—Stresses resulting from differences in elevation and loads were studied by D. H. Darwin.¹ The intensity of the stress in a series of uncompensated mountains and valleys varies with the height and density, but the depth of greatest stress depends wholly on the distance from crest to crest. That depth is 0.163 times the wave length, where the topographic wave length is the distance from crest to crest (Fig. 121).

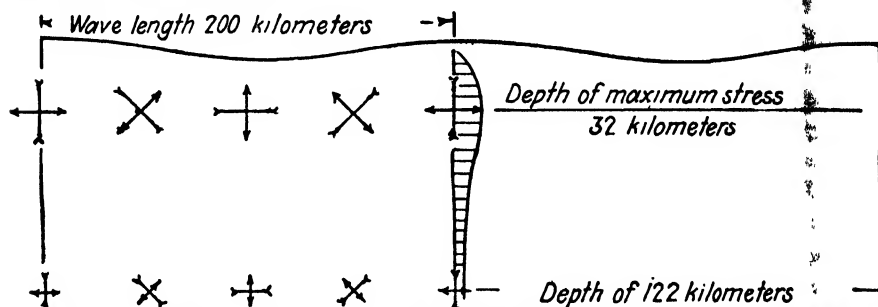


FIG. 121.—Diagram of the depth of maximum stress under loaded portions of the earth. (After Barrell, *Jour. Geology*, vol. 22. p. 734, 1914.) Arrows show the direction and magnitude of the stress differences under a series of uncompensated mountains and valleys. Mountain crests are drawn as 5 km. above valleys. With mountains 200 km. apart the maximum stress is at 32 km. below mean level.

This distribution of stresses is perhaps best realized by recalling the "slides" in the Culebra cut of the Panama Canal. These were not surface land slides approaching the "angle of rest," they were bulges in the bottom of the canal, resulting from deep-seated lateral thrust toward the canal from under the hills on each side.

Assumed topography	Distance apart, kilometers	Approximate difference in altitude, meters	Depth of maximum stress, kilometers
Mountains.....	12	2,000	2
Mountain ranges.....	180	3,000	30
Actual maximum (Pacific).....	400	4,000	64
Mountain systems.....	2,000	2,000	350
Continents.....	6,000	1,000	1,000

¹ Roy. Philos. Soc. Trans., vol. 173, pp. 187-230, 1832.

We are now in a position to calculate where the maximum stresses occur on account of actual topographic features.

The intensity of the stress is also estimated at not more than 37 per cent of the hydrostatic lateral pressure normal to such depths.

STRENGTHS OF ROCKS

Data. Tests of Strength.—Rocks have been tested under a variety of conditions. The pressure on rocks deep in the earth exerted by the overlying rock is much greater than that necessary to crush a cube of rock in a testing machine at the surface. The conditions at depth, however, differ from those in a testing machine in the fact that the mass under pressure is confined on all sides. Below a depth of $2\frac{1}{2}$ miles rocks will not crumble as at the surface but will flow, if the differential pressure is sufficient. The yielding of a rock at depth under such conditions does not weaken it in such a way that it yields much more easily to further deformation. Under a confining pressure equivalent to a depth of a few miles the differential pressure needed to deform a rock becomes very much greater than where unconfined. Adams's experiments show that at a depth of 22 miles it is *colossal*¹ (see Fig. 123 and page 389).

Most strength tests, however, are rapidly applied. It is known that even a small stress if long continued causes deformation, and there is a great volume of evidence that in the crust—especially near batholiths—rocks have been greatly deformed. This does not imply any fluidity, plasticity, or even conspicuous weakness. Tombstones and window caps may sag notably, but remain strong rocks.² It is wholly uncertain, however, whether or not cold rocks would yield in a few hundred years, as the tombstones do, if they were under a considerable confining pressure. Such data would require experiments prolonged through several lifetimes.

The flow of glacial ice has been much studied and is suggestive here. Ice flows by (1) idiomolecular exchange between crystals and internal gliding inside the crystals, (2) solid shearing of aggregates of grains, (3) thrust faults (both in the ice and between it and the rock). The shearing and thrusting are intermittent, but the recrystallization may be slow and continuous (page 402).

The effects of high temperatures on the strengths of rocks are not accurately known. Adams found a conspicuous temperature effect in limestones, and it is well known that metals and clays as well as ice are weakened by heat (page 390). It seems perfectly safe to assume that rocks are much weakened by heat, but the extent of weakening is very uncertain.

¹ ADAMS, F. D., and J. A. BANCROFT, Internal friction in rocks, Jour. Geology, vol. 25, pp. 634-635, 1917.

² WINSLOW, Flexibility of limestone, Am. Jour. Sci., 3d ser., vol. 43, pp. 133-134, 1892. Report on building stones, Tenth Census U. S., 1880, vol. 10, pp. 366-367.

Finally, we have a series of tests on glasses, supercooled liquids of almost infinite viscosity. A glass may be supercooled several hundred degrees (more than the cooling likely to have occurred in any masses at depths of a few hundred kilometers since the formation of the earliest known rocks), and still the glass is so low in viscosity that a stress of a few moments' duration will easily deform it.¹ There is little chance of a permanent magma layer in the earth, under the peneplaned mountains or the great river deltas that have not depressed their floors (see page 183 on zones or shells in the earth).

More indirect evidence of strength and weakness in the earth is based on its behavior to great stresses; earthquakes, body tides, rotational stresses, erosion and deposition, orogeny, and local bodies of rock with exceptional densities. For each of these we have some data, but it is necessary to refer to some theory in order to describe most of the data.

Astronomers studying the body tides in the earth, and its form, and the shifting of the earth's axis (data which can hardly be abstracted here) agree that the earth exhibits a general solidity and elasticity, and a rigidity twice that of steel. The *strength* however, may not be wholly determined by these tests. Most of the stresses are rapidly applied and released and do not prove that the earth would resist slowly applied long-continued stresses.

The earthquake records contribute a further mass of detail on the transmission and reflection of waves, but the data have been variously interpreted. It may be said that there are changes in the rate of transmission, indicating heavier rocks below the crust, and there are reflections of waves, indicating such a discontinuity of material as to suggest shells of different constitution. The rapidity with which earthquake shocks are applied and withdrawn gives but little data on the strength under long-continued stress.² Daly even argues that a liquid might transmit such shocks (see list of readings).

The geologic record furnishes abundant evidence of base leveling—times when a great load of rock is removed from high land and deposited in geosynclines. At other periods these geosynclines become crumpled into mountains. Geologists have for this reason been inclined to attribute to the rocks of the earth strength enough to support a very considerable load. The data are not exact.

On the other hand, geodesists have a considerable body of facts indicating that at most places on the earth there is no very great load of excess material. The deltas and high mountains are not excess masses supported wholly by strong rocks below, and the crust under a region of

¹ ADAMS, L. H., Temperatures at moderate depths within the earth, *Washington Acad. Sci. Jour.*, vol. 14, pp. 464-467, 1924.

² DALY, R. A., The outer shells of the earth, *Am. Jour. Sci.*, 5th ser., vol. 15, pp. 108-135, 1928.

base level is not relieved of so great a load as would seem from the removal of a mountain. The high areas on the surface of the earth seem to consist of rocks of relatively low density and the low areas of rocks with higher density. There are large volumes of interesting data, which may indicate the nature and strength of the material in the crust of the earth. Most of these are accurate observations on the attraction of gravity and on the deflection of the vertical.

Isostasy and Elastasy.—Isostasy is the condition of equilibrium of figure to which gravitation tends to reduce a planetary body. The earth not being homogeneous has a bulge where matter is light and a depression where it is heavy. The theories of isostasy (of which there are several) usually include the notion that at some depth below sea level (the so-called depth of compensation) all the upper parts of the earth rest on the mass below with equal pressures. This is probably not capable of very rigid proof. When erosion, sedimentation, volcanic action or mountain-making thrusts shift loads over the earth's surface, they disturb the equilibrium. The only way to maintain uniform pressures below is by a deep-seated adjustment or compensation. The facts seem to indicate some *approach* to such a state of balance, and it is common to refer to processes by which the earth draws toward this state as "isostatic adjustments." The uncertainties in the theories do not concern the reality of adjustments but rather their mechanism and location and completeness. (See the list of further readings.)

Chamberlin maintained that the "isostasy" proposed by Dutton¹ in 1889 implied to the geologists of that time the idea of blocks floating on a liquid, like the icebergs in a sea. If this is implied, the term should be abandoned, for the adjustments seem clearly to occur in an elastic solid. True isostatic adjustments above a liquid may possibly occur locally in the fault blocks forming the roofs about a mile thick over the great batholiths, some of which are more than 100 miles across. The ocean also is isostatic in the old sense.

Chamberlin² suggested the term "elastasy" for the equilibrium, and "elastatic" adjustments for the solid flow by reorganization, quite distinct from any fluidal flow. This term should not mislead anyone into thinking that all adjustments in the earth are by deformation "within the elastic limit." After deformation within such a limit, a deformed rock, if released, returns to its undeformed size and form. Rocks below the crust appear to have been deformed much beyond that limit. As noted later in the discussion of metamorphism (pages 390 and 391), deformation may be *clastic* if a rock breaks, or *plastic* if the crystal structure is deformed by gliding, bending, etc., or *blastic* if the material recrystallizes. The deformation involved in

¹ DUTTON, C. E., On some of the greater problems of physical geology, *Philos. Soc. Bull.*, vol. 11, pp. 51-64; reprinted in *Washington Acad. Sci. Jour.*, vol. 15, pp. 359-369, 1925.

² CHAMBERLIN, T. C., Intragology—elastasy vs. isostasy, *Jour. Geology*, vol. 35, pp. 89-94, 1927.

adjustments of load in the earth are no doubt blastic, and this should be remembered whether they are called isostatic or elastic.¹

Geodetic Data.—Gravity data are obtained by accurate observations of the attraction of the earth at various stations. At first thought it might seem that the earth, having a constant mass would have a uniform gravitative attraction, but gravity varies with both mass and distance. High-mountain stations are farther from the center of gravity of the earth than sea-level stations. The mass of the mountain is a mass above and added to that below the mean level of the earth, and adds to the attraction

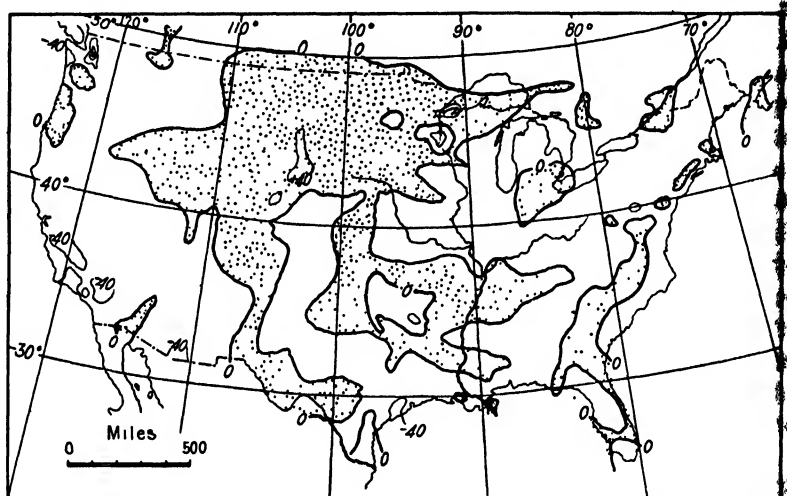


FIG. 122.—Map of the United States showing by contours the gravity anomalies. (After Bowie.) Stippled areas positive anomaly; plain areas negative anomaly.

locally. When the effects of elevation of the station and of the inequalities of surface are calculated, and the observed gravity is properly corrected, it is found that corrected values of gravity still vary. The differences from mean value are called "gravity anomalies," and are used as a basis of calculation of the strength of rocks and nature of material below (see Figs. 122 and 123).

Data as to the vertical are obtained by comparing the direction indicated by a plumb bob, which is supposed to hang pointing toward the center of the earth, with astronomic observations as to the direction of the center of the earth. It is easy to see that if a sensitive plumb bob was hung so that a mountain mass lay on one side and no corresponding mountain lay on the other side, the mountain would draw the bob over a little toward itself. An interesting fact about observations near mountains is that although a plumb bob is pulled over toward the mountain, it is not

¹ Blastasy might be a better term than either isostasy or elastasy, referring to the approximate equilibrium attained by blastic deformation or adjustments in an elastic solid.

deflected so much as would be expected from the size and material of the mountain. It is clear that beneath the mountain there must be rocks that are light and attract the plumb bob less than the rocks of the low land on the other side of the plumb bob.

The observations of gravity and of the deflection of the vertical are more characteristically numerical and precise than geologic records of erosion, deposition and mountain building, and they have been made the basis of very elaborate calculation. Much of the calculation, however, involves further assumption, and the conclusions can not be classed as facts.¹ Using such instrumental exactness, the geodesists have impressed many geologists with the lack of exactness in geologic data. Nevertheless the geologic data can not be ignored, for several have a definite bearing on the theory.

Geologic Data.—Rocks are clearly not strong enough to support the continents as a flat arch. Yet the present continents seem to have been continental areas in general throughout geologic ages, showing that uplift has roughly kept pace with erosion. In certain areas, however, it is found that there has been a gradual settling to depths of more than 30,000 feet, proved by the continuous deposition of shallow-sea sediments to such thicknesses. The depth of some coral sediments also shows progressive settling. There are faults with a throw of many thousand feet, and the absence of any evidence of such differences in elevation shows that the faulting was gradual and progressive. The blocklike form of continents and ocean basins gives us another suggestion of vertical movement of blocks of the crust. Paleontology strongly indicates that continents once existed where there are ocean basins now. The height of the uplift after the Pleistocene glaciation is shown by the elevations of glacial-lake beaches and corresponds fairly well with what might be expected to restore equilibrium after the melting of a continental glacier. Finally, the densities of igneous rocks in high areas are less than those in low areas.

On the other hand, although both the geodetic and this much of the geologic evidence indicate an adjustment to a certain degree of compensation, geologic evidence sets some very definite limits. Volcanic mountains grow and only cease to grow from the hydrostatic pressure of a lava column; they do not cease to rise higher because of any weakness of the floor. Again, if adjustment was complete, it should occur gradually and constantly, but geologically we find great deformations that are catastrophic and, at least locally, periodic.

Theory. Estimates of Strength.—Most of the data on the strengths of rocks are on rocks unconfined and with the pressure rapidly applied, but a few observations have been made on rocks subjected to stresses for several hundred years. There are apparently no data at all as to the behavior of rocks under prolonged stresses with confining pressures.

¹ BOWIE, WILLIAM, *Isostatic investigations and data for gravity stations in the United States*, U. S. Coast and Geodetic Survey, Special Pub. 99, 1924.

We may perhaps be safe in estimating the strengths of rocks under stresses of a few hours' duration to depths of 30 or 40 km. Below that, and for prolonged stresses at *any* depth, we can only infer the strength from the behavior of the mass under the greater geologic forces. At depths around 20 miles the strength of rocks under rapidly applied stresses is colossal. It may be that during orogenic revolutions the forces applied are also colossal, but it seems very questionable whether the forces need be so great; more likely they are prolonged, and under such forces the rocks may not have such great strengths. The geodetic data show that if the stresses are equivalent to a load of several thousand feet over a large area, there will be some adjustment. The conflict between the laboratory evidence of colossal strength and the geodetic evidence of weakness is explained by petrographic studies of rocks. *Rocks yield by recrystallization.*

Willis draws an interesting comparison between the pressure of rock load at depth and the probable strength at such depths, concluding that below 40 miles the rocks would fail unless supported by lateral pressure, even if the strength continued to increase as shown in Adams's experiments.¹

How large a stress serves to deform rocks at depth? The chief studies of this matter were made by Barrell, and it is well to have some familiarity with his terminology.² Having studied the stresses that are produced by a shift in load at the earth's surface and estimated from the anomalies of gravity how great an excess of material might remain uncompensated, he tabulated his estimates of strengths at various depths. The estimated curve of strength rises rapidly to depths of 30 km. and falls off even more rapidly below that. At 50 km. Barrell thought that rocks might be only 25 per cent as strong as at the surface and at 400 km. only 4 per cent as strong as at the surface. The depth zone where rocks are so weak he called the "asthenosphere" or sphere of weakness.

Barrell's estimates refer little, if at all, to the duration of the stresses, but his discussion shows that he had in mind such stresses as those produced by erosion—stresses that accumulate gradually and are applied for thousands of years. As the duration of the stress may make a great difference, it may be permissible to estimate roughly the strengths under three sets of conditions.

Figure 123 shows the stresses due to rock loads estimated from a study of gravity anomalies; and the depths of maximum stresses estimated as in the discussion of density (page 160). Interest centers in those excesses and deficiencies that show a maximum intensity of stress at moderate depths, because the deepest stress, due to differences in the

¹ WILLIS, BAILEY, and ROBIN WILLIS, "Geologic Structures," 2d ed., p. 444, McGraw-Hill Book Company, Inc., 1929.

² BARRELL, JOSEPH, The strength of the earth's crust, Jour. Geology, vols. 22 and 23, especially vol. 23, pp. 27-44. See also BARRELL, The spheres of the earth and their effects on the lithosphere, Am. Jour. Sci., 5th ser., vol. 10, pp. 505-529, 1925.

material of continents and ocean basins, is of small intensity. Note that these maximum stresses are much less than the strengths of rocks at the surface as found in testing machines, but these are the loads that in general seem to be largely compensated. In other words, if a local area is loaded with a mountain range, the rocks yield.

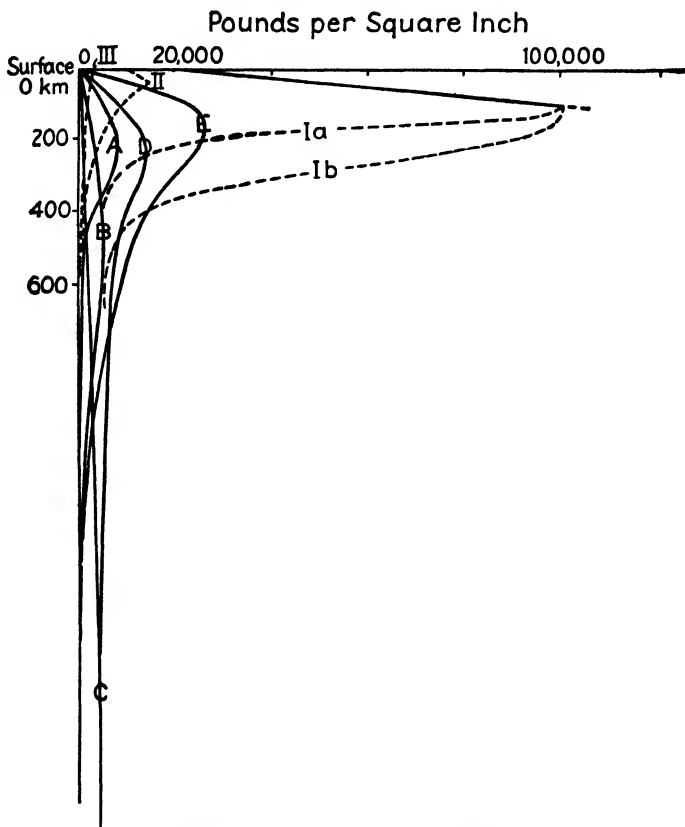


FIG. 123.—Probable stresses (lettered curves) and estimates of strengths of rocks (curves with Roman numerals) at various depths in the earth.

Curves of stresses *A*, *B* and *C* result from masses in the earth that the geodetic data indicate are uncompensated. Curve *D* is the sum of *A*, *B* and *C*. Curve *E* shows the stresses that might arise if the earth's topographic features were uncompensated.

Curves of estimated strength. *I*, for rapidly applied stresses. (After Adams and Bancroft, *Jour. Geology*, vol. 25, pp. 597-637, 1917.) *Ia* based on geodetic data and *Ib* based on geologic data. (After Barrell, *Jour. Geology*, vol. 23, p. 33, 1915.) *II*, for stresses slowly applied for a few years. *III*, for stresses very slowly applied during hundreds of thousands of years.

On the other hand, the "gravity anomalies" found by the geodesists can not all be explained by the assumption of light rocks under mountains and heavy rocks under basins. If a calculation is made on the assumption of such differences in rocks, and a correction for this is applied to a map of anomalies, about nine-tenths of the anomaly disappears. The other

one-tenth is probably attributable to the local excesses or deficiencies of masses that are supported by the strengths of rocks. The masses supported are variously estimated as equivalent to rock 1,000 to 2,000 feet thick over the area of a state.¹ A much greater thickness might be supported if not spread over so wide an area.

Depth of Compensation.—Assuming as seems reasonable that the earth approaches more or less closely to a condition of equilibrium, there are wide differences of opinion as to the depth at which adjustments may take place in the earth. The depth is probably much too uncertain and variable to be described as a “level of compensation.” In the first place it is probable that the compensation is only partial; in the second place it probably varies in depth from time to time and from place to place; no doubt some adjustments occur at a depth of a mile or two, and others may involve very great depths. By varying the assumptions the calculated “level of compensation” may be shifted from near the surface to depths of several hundred miles. (See the list of further readings.)

The geodesists believe that the conditions all over the earth favor isostatic adjustments at depths of comparatively few miles. Their estimates of the level of compensation, however, are much more precise than the indications from geologic evidence of various conditions in different parts of the earth. The estimate of 76 miles below sea level has been so widely accepted and used that it may have greatly delayed our progress toward a correct understanding of the interior. The sequence in their estimates is also somewhat instructive: 113, 122, 95, and finally 60 km. Probably the conclusion most generally agreed to by geodesists is that gravity anomalies are local effects from rock masses within 100 km. of the surface.

Several geodesists now frankly admit that (1) there is no sharply defined depth of compensation, (2) it varies in different regions, (3) the density differences are not uniformly distributed vertically as assumed in the geodesists' calculations. No geologist should be misled by the definiteness of the results of their calculations, into considering their assumptions are thereby proved correct.

Dr. T. C. Chamberlin advocated compensation by balancing sectors in which compensation extends down toward the center of the earth perhaps more than 1,000 miles. The arguments for this idea have not been published with mathematical detail, as the geodetic data have been. For the present, then, the zone in which adjustments occur normally and universally may be supposed to extend from approximately 60 to 600 km. in depth.

How Deep-seated Adjustment Is Accomplished.—The idea of an undertow near the surface has little to support it. The reaction is deep, unless the conditions normal at depth are locally and temporarily brought up closer to the surface by igneous action.

The blocks have thus far been considered as if independently movable. Of course they are attached to their neighbors and adjustment is by rock

¹ BARRELL, *op. cit.* SWANSON, *Jour. Geology*, vol. 36, pp. 416–429, 1928, referring to COLEMAN, *Geol. Soc. America Bull.*, vol. 31, pp. 326–327, estimates 1,000 feet thickness over an area 200 miles wide. T. C. CHAMBERLIN, *Jour. Geology*, vol. 34, pp. 1–28, estimates 1,600 feet. Hayford and Bowie have made rather smaller estimates, down to an area of 1 square mile.

flowage and tilting, unless, as in some regions, there are block faults. Distributed faulting and "punching" of blocks are entirely possible, but underneath the blocks the rocks flow.

Suppose erosion of $\frac{1}{2}$ km. of the high area with deposition of the load on the low area, and suppose this to be adjusted by undertow below. The zone of adjustment no doubt grades into the zones above and below, but is probably five or ten times as thick as the lithosphere.¹ The nature of the undertow when compensation occurs is shown by the curve (Fig. 124). The motion from *B* to *A* may crowd the point *E* possibly 2 km. to the left of its original position. As this motion fades out to zero at the top, the differential motion across the line *DF* is perhaps 10 meters per kilometer. That is, in the adjustment any particle on the line *DE* may

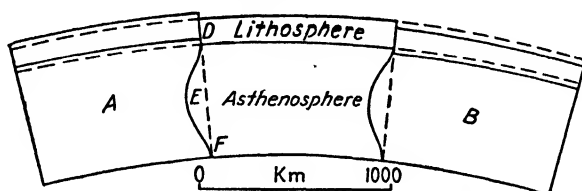


FIG. 124.—Diagram of the nature of the undertow necessary to restore equilibrium after erosion and deposition. (After Barrell, *Jour. Geology*, vol. 22, p. 681, 1914.) Topography greatly exaggerated.

move about a meter more than a particle 100 meters higher up. Clearly the deformation is not intense.

If a mountain 1 km. high is thus eroded and adjusted by an undertow, the densities of surface rock transferred and deep rock undertow probably differ; they are perhaps 2.7 to 2.9 in the surface rock and 3.2 to 3.4 at depth. The rise of the mountain by adjustment will not bring it as high as it was at first. A second erosion of the mass raised will be adjusted with a still smaller rise of the surface. The process "runs down." It should be noted, however, that some 6 or 7 km. would thus be eroded before a mountain 1 km. high would be base leveled with complete adjustment. The base leveling of some high mountains would by this process require more erosion than is believed to have occurred in all geologic time. As we know that a mountain hardly persists through a single period, to say nothing of an era, it is not likely that base leveling involves complete compensation through the whole process.

The Asthenosphere.—The basis for the assumption of an asthenosphere is the combination of two facts: (1) the increase of strength shown by rocks under confining pressure from some 20,000 pounds per square inch at one atmosphere to several times that strength at pressures equivalent to 20 or 30 miles of depth; and (2) the geodetic evidence that rocks have yielded to the stresses created by erosion and deposition of mountains—very moderate stresses, which would not produce rapid deformation of rocks in a testing machine, unconfined.

¹ DeLury, J. S., The auto-traction hypothesis of crustal evolution, contribution from the department of geology and mineralogy, University of Manitoba, 1931; suggests that motion, once started, is localized by the heat of friction.

Rocks yield and are deformed by several distinct processes, sometimes called twinning, gliding, strain, granulation, shearing, and recrystallization. See the section on metamorphism. In the deeper parts of the earth no doubt recrystallization dominates in all deformation. The conditions favorable to recrystallization are (1) heat, (2) water, (3) non-uniform pressure, (4) long time, (5) weak rocks. Of these factors, different ones are emphasized by different geologists. It seems almost certain that heat dominates at depths of 100 to 600 km., and if so the assumption of an asthenosphere is not unreasonable.

Chamberlin,¹ noting that slow, long continued stress will deform even a slab of rock at the surface within historic time, maintains that rocks will yield in any part of the earth where the stresses are applied slowly for a long time. No zone is recognized as having conditions distinct from its neighbors. He assumes that rocks under confining pressure yield as they do at the surface, but for this assumption we have no data.

Leith² notes that so far as we see the results of rock deformation, the upper zone of fracture gives place in depth to a zone of *combined fracture and flowage*. This second zone extends as far as any rocks exposed at the surface have ever been buried. He, therefore, questions any extrapolation to a zone where rocks flow without fracture though at the greater depths flowage increasingly dominates (see page 403). If we start with the known batholiths of any period and insist on extrapolating their known forms, widening in depth, each would soon join its neighboring batholiths in a continuous layer of granitic magma. Between the idea of a magmatic layer and the idea of a zone of combined fracture and flow, both of which are rigid and strict extrapolations, an asthenosphere or zone of easy flow seems to be a very logical compromise.

Holmes³ presents another objection to the idea of an asthenosphere, based on the radium content of rocks and the thermal history of the earth. As Barrell seemed to conclude that the asthenosphere generates basaltic magma, Holmes assumed that the asthenosphere must have the average radium content of basalt and so evolve more heat than it seems to have evolved, judging by the historical geology of igneous action. The question may yet be asked whether the basaltic magma rising from a great depth is proof of the basaltic composition of the asthenosphere. If the fusion is selective, the most fusible minerals and mixtures melting first, and if the small bodies of magma so formed rise as small units coalescing at the base of the lithosphere into reservoirs of considerable size, then it may still be possible that basaltic magma is generated in an asthenosphere of peridotitic composition.

It may also be questioned whether the data have proved that heat will be accumulated in the basaltic zone as fast as indicated by the radium content of surface basalts. Possibly some other reactions absorb heat in a way not now understood.

Before our consideration of the asthenosphere is closed, a more general statement is needed as to rock flowage. The discussion, up to this point, might imply that rock flowage by recrystallization is normally limited to rocks 60 km. or more down from the surface—at such depths

¹ CHAMBERLIN, R. T., Review of Bowie on Isostasy, Jour. Geology, vol. 36, pp. 181-184, 1928.

² LEITH, C. K., Structural failure of the lithosphere, Science, New ser., vol. 53, pp. 195-208, 1921.

³ HOLMES, ARTHUR, Radioactivity and the earth's thermal history, Geol. Mag., vol. 62, pp. 504-512, 1925.

that the rocks that flow would never be exposed to the light of day by erosion. No such impression is intended. The asthenosphere is a matter of *conditions*, mostly a matter of high temperature, and water. It will probably be agreed that rock flowage has occurred in some places so near the surface that later erosion has exposed the rocks. That is, the depth is not so great as that which rocks in other places withstand without sign of much rock flow. The question of the asthenosphere is whether the conditions for rapid rock flowage become quite general all around the earth at moderate depths, as distinct from the local conditions for such flowage, known to have been developed near the surface.

Possibly the best answer is that based on a comparison of the curve of temperature with the fusion-pressure curves. Rock flowage is favored when the temperature approaches the temperature of fusion at the prevailing pressure. It will be noted that the curves in Figs. 118 and 124 suggest that the conditions favoring rock flowage prevail at depths of 60 to 600 km. Such curves may be taken as a rough approximation to the general average of the earth, but both curves probably vary in different parts of the earth. Even the observed geothermal gradient and the observed rock densities producing pressures vary in different regions.

Locally where rocks have little radium, where circulating water cools the crust, where deformation has been slight and other factors contribute to the strength of the rocks, the zone of easy flowage may be greatly depressed.

At other localities, in the absence of these factors and especially in the presence of rising batholithic magmas, the conditions favoring rock flowage may rise to points within a mile of the surface of the earth. Local adjustments should be possible in the roofs a few miles thick over batholiths that spread over many miles laterally. Batholithic roof rocks injected, recrystallized, deformed and almost assimilated, but now exposed by erosion, give us our best clue to conditions that prevail widely in depths beyond all direct exploration. The rocks seem to have been weak enough to justify the assumption of an asthenosphere where such conditions become general all around the earth.

One of the districts where the depth and temperature effects seem to have been intense, is that north of Georgian Bay of Lake Huron, Ontario.¹ The peculiar features of a rock series that may indicate the condition of an asthenosphere are here listed, though some of these features may prove to be local rather than general.

1. Gradations are common from sediments to gneisses that appear to be largely igneous.

2. The apparently igneous gneisses occupy such a position in a structural belt and retain such internal structures as to indicate that they replace the sediments.

3. Igneous sheets are intruded in great numbers conforming to the bedded structure.

¹ COLLINS, W. H., and T. T. QUIRKE, The disappearance of the Huronian, Geological Survey of Canada Mem. 160, 1930. See also G. W. BAIN, Flowage folding, *Am. Jour. Sci.*, 5th ser. vol. 22, pp. 524-529, 1931.

4. The folds are commonly fan shaped (Fig. 240) or radial—more pinched than is common in regional dynamic metamorphism.

5. Certain igneous features, such as aplites and fish-shaped schlieren, are almost absent.

6. There has been notable deformation and dynamic metamorphism since most of the igneous material came to place.

7. The cleavage follows the bedding in all folds, in contrast to the well-known discordance in the dynamically metamorphosed rocks.

STRUCTURES INDICATING DEEP CONDITIONS

Data.—The surface of the earth may be very roughly divided into six "shield areas" and intervening basins, with a "hinge zone" as a narrow belt along the contact, where orogenic movement is active in the later geologic epochs. The structures suggest that at some early epoch the mobility characteristic of the hinge zone prevailed, though perhaps not simultaneously, in all parts of the shields. It seems likely that the shield areas, resistant to mountain folding, have grown by additions along their edges (Fig. 105).

The records of diastrophism indicate that troughs in narrow belts on the borders of shields, slowly deepen and are filled with sediment, the floors of the trough sinking gradually many thousands of feet. Then, suddenly in comparison with the period of sedimentation, the sediments are folded and uplifted. This change from sinking to rising implies some fundamental reversal of forces or processes. Large intrusives commonly follow the folding near the center of the structure. The largest batholiths are situated along the mountain systems, opposite the greatest overthrusts¹ but not in the overthrust slices.

Along the flanks of great mountain chains the dips of faults and the axial planes of folds and schistosity suggest that many mountain ranges consist of a wedge of material tapering downward. Laboratory experiments also may be offered as data on the formation of wedgelike structures.²

Theory.—Only a relatively small set of facts about the formation of mountain folds being available, the discussion is highly speculative.

The structures clearly show that some lateral force has thrust one portion of the crust against another. The nature of that thrust is very uncertain. The classic view is that a shrinking core of the earth causes lateral thrusts as the crust tends to collapse on the core. The shrinking of the core may be from cooling or from atomic changes (page 159). Neither cooling nor normal compression seems wholly adequate to explain the sinking of some of the great geosynclines. Perhaps, less clear, there

¹ KEITH, ARTHUR, Outline of Appalachian structure, Geol. Soc. America Bull., vol. 34, pp. 309-387, 1923.

² CHAMBERLIN, R. T., Appalachian folds of central Pennsylvania, Jour. Geology, vol. 18, pp. 228-251, 1910; The wedge theory of diastrophism, Jour. Geology, vol. 33, pp. 755-791, 1925.

may be thrusts in the crust as a result of deep-seated adjustments; or from changes of temperature in the crust itself; or from aggressive igneous intrusion. The late period of invasion of magmas, as compared with periods of folding, however, makes it doubtful if magmatic thrusts are usually the causes of folding.¹ A radically different theory of mountain folding suggests that a block of the crust drifts on a magma layer below, until it drifts against a resistant block of crust with force enough to crumple the edge. The drifting block may be moved by gravitative pull down an incline or by some force involving the earth's rotation and tidal attractions.² Still another theory suggests that the earth flows under

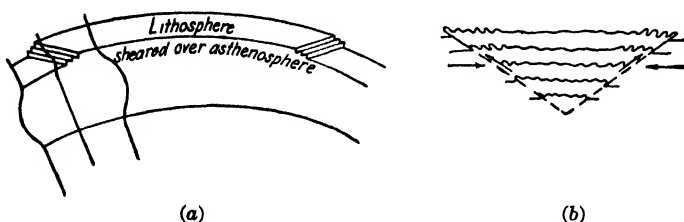


FIG. 125.—The structures supposed to exist below folded mountains.
(a) (After Barrell.) (b) (After R. T. Chamberlin.)

isostatic excess loads much as glacial ice flows, and causes folds in some regions and tension in others.³

Given the lateral forces capable of folding rocks into mountains, there is considerable disagreement as to the depth of material involved in the structure seen at the surface. The thrusts at the borders of the mountains, for example, are variously interpreted. If they persist downward from the observed faults, folds and cleavage, as planes or zones of movement, the wedges may be several hundred miles deep. Calculations of the depths affected by folding, based on uplift and apparent crustal shortening, also indicate great depths. But these are so much modified and by so many factors, that the deeper structure is not safely inferred. Two contrasting notions are shown in Fig. 125.

Any lateral thrust capable of crumpling rocks into mountain folds would almost necessarily shift material enough to disturb the earth's equilibrium. Yet gravity measurements indicate that mountains are not far out of adjustment. After the folding therefore, if not during its progress, the loads become largely adjusted, and the height of the mountains is to be attributed not to the folds but to the lighter material in the mountain areas.

The light material responsible for the elevation of mountains is variously postulated as sediments, as granite, and as chambers of magma, somehow concentrated in the

¹ BARRELL, JOSEPH, Relations of subjacent igneous invasion to regional metamorphism, *Am. Jour. Sci.*, 5th ser., vol. 1, pp. 263-267, 1921.

² "A Symposium on the Theory of Continental Drift," *Am. Assoc. Petroleum Geologists*, London, Murby and Co., 1928.

³ DELURY, J. S., The auto-traction hypothesis of crustal evolution; contribution from the department of geology and mineralogy, University of Manitoba, 1931.

area. The formation of magma may readily be understood as an explanation of the presence of light material, for a magma occupies more volume than crystalline rock of the same composition. The accumulation of sediment in a geosyncline and its concentration in a small area by crumpling is also an easily understood explanation for large volumes of light rock. Neither of these, however, satisfactorily explains the mountains old enough to have their magmas solidified and their sediments mostly eroded. These masses seem to be very largely granite, a rock lighter than the average igneous rock, and they still stand high probably because they are composed of this light rock. The processes by which the concentration of granite in certain regions was brought about are not well understood, especially as some of the concentration occurred early in the earth's history, forming light masses, which have persisted ever since as continents.

DEEPER PARTS AND SOURCES OF BATHOLITHS

Data.—Under many extensive sediments the drill has found igneous rocks, and it is believed that igneous rocks underlie all series of sediments. The structure of the great "shield areas" of the earth indicates that even

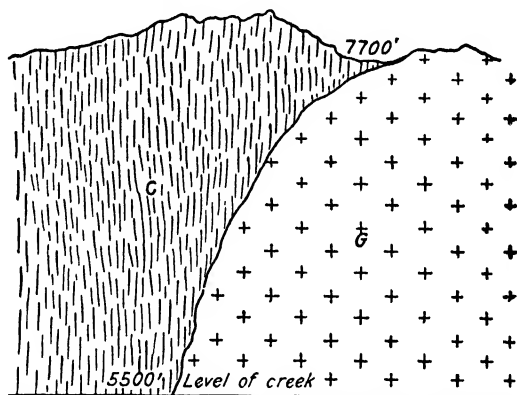


FIG. 126.—Sketch of Canyon wall north of creek at Castle Peak, British Columbia. (After Daly.)

the deepest of the residual patches of sediments have been intruded by granite from below.

These deep granites and other great masses of granite are commonly called batholiths. The floors or roots of the large masses are unknown, but certain observations indicate the average nature of such bodies. Excluding the few masses of very moderate area that have visible floors, these granites appear to be larger in horizontal extent the greater the depth for many thousands of feet (Figs. 126 and 6). There are satellitic masses near some large batholiths, which are so clearly related that the underground connection is very certain, and the connection of a few has been proved by exploration (Fig. 127). The metamorphism in the roof rocks of batholiths and those of associated small intrusives is very characteristic, and exactly similar alterations are known in several large areas

where batholiths are not exposed, so that the probability of a batholith at no great depth below is very strong. Structural work showing that certain small granite intrusives have floors does not in any way affect this conclusion from large batholiths.

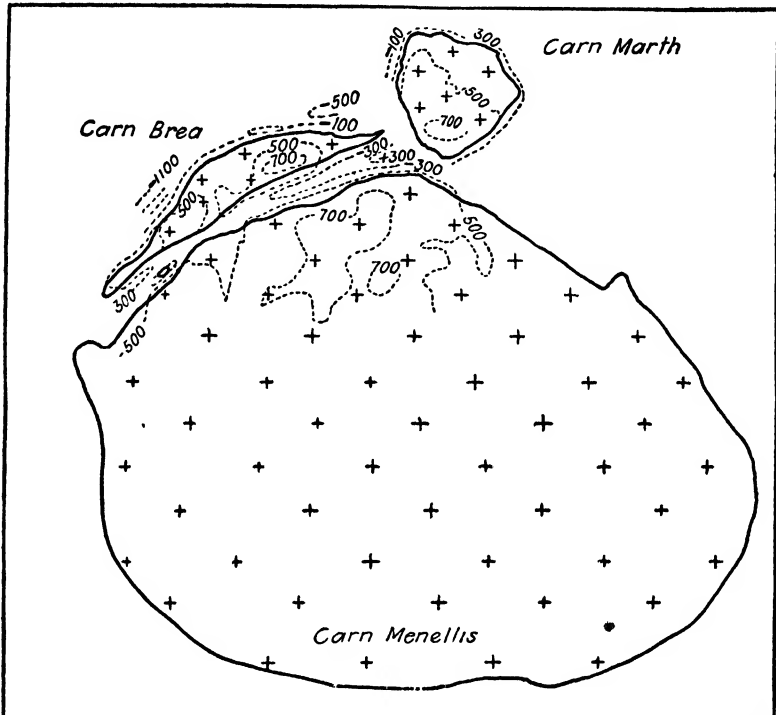


FIG. 127.—Sketch map of Carn Menellis and some near-by granites that rose through sediments of Cornwall. (After Hill, MacAlester and Flett.) The outcrops of granite are enclosed in the heavy lines and the surface of the granite is indicated by the dotted contours at certain levels above or below sea level. The underground connection of the masses that form three distinct surface areas is clearly proved by mining explorations. Three miles to one inch.

Most batholiths lie in orogenic belts, as is strikingly shown in the geologic map of North America. They are elongated with the belts and centrally placed as regards the area of folding. Although some batholiths lie in plateaus, the relation to mountains is too general to be accidental. As noted on page 172, the greatest masses of igneous rocks in the Appalachians are opposite the greatest overthrusts, but not in the overthrust slices. Commonly the magma came to its place in the orogenic period, but some deformation may have followed or alternated with magmatic invasion. It is clear that there is a relation between batholiths and deformation, but we may still be uncertain as to which is cause and which is effect, especially as some of the batholiths in plateau areas may be intruded in rocks that are almost wholly undeformed (pages 199 to 204).

The roofs of batholiths are low domed or flat, and their sides are steep. At lower levels the walls are schistose, as though schist was crowded aside. In composition batholiths are mostly granitic varying to granodioritic and locally including phases as basic as gabbro. It is not possible to determine what lies below the granite of most batholiths, even where they show basic phases in parts of the exposures. The largest rock masses exposed from top to bottom, however—the lopoliths and large sills—are differentiated from granitic tops through large bodies of basaltic composition to a base rich in olivine (pages 132 and 133). The granite of batholiths forms a top phase, which is still larger than that of lopoliths.

Two other kinds of data already noted may have some bearing on the estimates of the nature of batholiths namely, the signs of crustal shortening (page 159), and the fact that there is commonly a cycle of igneous action related to orogeny (page 207). It has also been noted that different igneous rocks differ in the average content of radioactive minerals producing heat (page 147) and that all rocks are poor conductors of heat.

Theory. Roots of Batholiths. *Composition.*—A number of geologists in discussing batholiths assume, either tacitly or verbally, that a granite batholith must have developed from a magma that was essentially granitic in composition. They assume that the earlier history of the magma is not safely inferred. The evidence, however, is becoming stronger every year for the theory that a large proportion of primary magmas are basaltic to dioritic and may have even more basic phases below (see page 190 *et seq.*). The known granites are becoming in theory more and more closely related with large bodies of basic rocks, and the granites prove to be relatively small upper phases of the masses. Some small granites with visible boundaries and no basic phases, may be satellitic.

Form.—Daly has emphasized the fact that the bottoms of batholiths are unknown, and he uses the term "subjacent," to distinguish these masses from "injected" igneous rocks of other forms. The distinction is very significant, and indicates a belief that subjacent bodies have wide, direct, open connection with the deep supplies, sources or chambers of magma. The origin and mode of emplacement of batholiths are still matters of controversy. Batholiths may rise by assimilation, by thrust and lift, by stoping and replacement, or by crowding aside their walls. Seven rough sketches (Fig. 128) may serve to illustrate the wide divergence of opinion as to the conditions below the granite exposures.

The idea of extensive rise by assimilation has little support (see page 228).

Three lines of evidence may be cited as opposed to the idea that most batholiths have floors at moderate depths, much like laccoliths or sills,

also opposed to the notion that they are "tack shaped" with only dike-like feeders: (a) The petrographic detail shows in many large batholiths a differentiation from basaltic to granitic nature, and almost everywhere that differentiated masses are well exposed as large intrusives, the granite makes up only about 10 per cent of the mass. If the granite portion is several thousand feet thick, any floor that the mass may have is probably more than 20 miles down. (b) Ore-vein deposits are rarely found related to masses with known floors and only to huge masses; all over the world

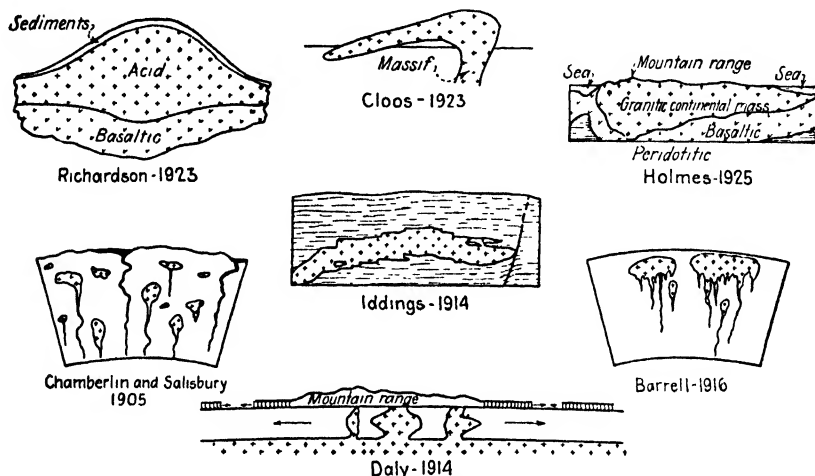


FIG. 128.—Sketches indicating the several suggestions as to the deeper structure under granite masses in the earth.

ore veins are characteristically related to granitic masses for which floors, if any, are so deep as to be unknown and probably more like roots than like the floor of a sill. (c) The structure shown by the roof and walls in both maps and cross sections is so clearly transgressive in some plateau batholiths, that the evidence of extensive stoping is conclusive.

The open connection of batholiths with their deep seated reservoirs or sources is therefore favored, in spite of an occasional occurrence of granite apophyses with known floors. (See the list of further readings.)

It is quite possible that differences in theories as to batholiths are partly attributable to differences in the ideas of deeper sources of magma. If there are enormous reservoirs of magma at depths so great that they are never exposed, or if there is a fused zone permanently or periodically in the earth, then the batholiths may logically be considered as being fed from the deep sources through either a thrust opening or a dike fissure, but the weight of evidence is against any such magma zone (pages 182 to 184). On the other hand, there is very clear evidence that any large deep-seated chamber full of magma would tend to rise, either by distending its wall and roof or by stoping. These processes mean that the magma in depth evolves from that deep reservoir into such masses as the known batholiths. If the accumulation of the deep reservoir was under a plateau region it might form a plateau batholith. But the

chances of its formation are better under the region of mountain building. The friction of active orogeny may be a large factor in magma formation.

THE CASE FOR ZONES OR SHELLS IN THE EARTH

The following are the zones or shells that are known and those that have been suggested:

Known shells	{	The atmosphere	{	Lithosphere	
		The discontinuous hydrosphere			
		The discontinuous sedimentary shell			
		The probably discontinuous granite shell			
		The probably continuous basaltic shell			
Inferred shells	{	The probably continuous peridotitic shell	{	Asthenosphere	
		(Gradation)			
		Largely metallic core, or planetesimal nucleus; this also may be zoned			
				Akthosphere	{
				Barysphere	
					Centrosphere

The known shells require no discussion here.

Data.—Many of the data cited above apply to the discussion of the theories about zones: The facts that the average rock at the surface is quartz-monzonite (page 128), that the density is much greater for the earth than for the surface (page 157), that the large igneous intrusions are gravitatively differentiated (pages 132 to 136), that there is local, if not general, periodicity in base leveling and diastrophism (page 159), and that the strengths of rocks seem to vary with depth (pages 161 to 167). A few other items, which have more significance here than elsewhere, will be added.

The universal prominence of basalt in all parts of the world, and throughout geologic time, warrants a belief in a continuous mass of material of basaltic composition underlying the more-silicic rocks observed in many places. There is no certainty that this mass is uniform in thickness or distribution or that it has the texture or minerals of basalt. Rocks as different as eclogite have nearly a basaltic composition.

The earth is magnetic, and this fact suggests to many men that the earth's core is of iron. Although high temperatures may affect the magnetism of iron, the alloys of iron may not be so much affected, and high pressures may counteract the influence of temperature. Furthermore we are not at all certain how hot the center of the earth is. The meteoritic material reaching the earth from outer space has about equal amounts of metallic and stony material, but estimates are very rough.

Seismologists have a large volume of data on the transmission and reflection of earthquake waves, some from carefully controlled shocks produced by great explosions.¹ It is well understood that the speeds of waves will vary with the densities of the rocks, and if a wave passes obliquely from one rock to another, it may be partly refracted and partly

¹ WRINCH and JEFFREYS, Roy. Astron. Soc. Monthly Notices, Geophys. suppl. 1, pp. 15-22, 1923.

reflected. A calculation of the speed of earthquake waves to be expected from the known compressibility and density of the rocks shows that the fast wave should have a velocity of about 5.6 km. per second in granite and about 7.9 km. per second in eclogite or peridotite. From the seismographic records of an explosion at the surface there were found fast waves with a velocity of 5.4 km. per second, agreeing very well with the idea of a granite shell nearer the surface than the heavier rocks, at least over continental areas. The observed velocities not far below the surface are generally about 7.9 km. per second, suggesting that material of such density as eclogite or peridotite may occur at comparatively shallow depths. There seems to be general agreement that some waves are reflected, indicating discontinuities of material in the earth, perhaps at several horizons, also that the depth of one of the reflecting surfaces is greater under continents than under ocean basins. (See list of further readings.)

The folded and thrust structures seen at the surface (page 173) suggest wedges extending to such depths as to transgress the structure of any supposed upper shell. The upper parts of the wedge structure are clear, but the deeper parts are only inferred.

Theory.—The thicknesses of granite masses are variously estimated from none under the ocean basins to 30 or 40 km. under the mountains of the continents. If this material gives way at depth to material of basaltic composition, somewhat abruptly, this fact may explain the known reflection of earthquake waves. Such an abrupt change is in fair agreement with the make-up of large differentiated intrusives, which show granite above gabbro with a very thin gradation zone. The depth of the base of the granite, however, and the depth of any discontinuity indicated by earthquakes,¹ are so variable as to make it very questionable if the granite makes a zone or shell in any ordinary sense. There is much doubt also whether the discontinuities are really sharp. The lower parts of granites are probably at very different depths in different places even in one continent or in one mountain range.

The supposed basaltic shell though more wide spread is deeper than the granite and less accessible, and the earthquake data become less easily interpreted as the waves reach greater depths. Some estimates place the basaltic compositions at 0 to 8 km., others as deep as 60 to 600 km. Perhaps it can be agreed that the speed of waves at 60 km. indicates a density about that of peridotite, but beyond that there is little agreement. Peridotite is only one of several possible rocks of similar density. Eclogite, with a composition much like that of basalt, is equally dense. The known masses of peridotite intrusives have few inclusions of basalt, and this fact casts some doubt on the reality of a basalt zone above

¹ MACELWANE, J. B., The interior of the earth, *Seismol. Soc. America Bull.*, vol. 14, pp. 81–87, 1924; suggests several zones but not much accuracy as to their depth.

peridotite.¹ Other rocks of about the composition of basalt may well supply basaltic magma.

Differences in the speeds of surface waves on continents and under ocean basins have led to the suggestion that the granite zone is absent under oceans, and that the basalt zone of the ocean connects with a zone of rock below the granites of the continents.

Below the basaltic material, it has been suggested, peridotitic material may constitute an immense shell. The bases of the suggestion are the density indicated by the speed of earthquake waves, the known differentiation of large basaltic magmas to a basal peridotite and the abundance of olivine in meteorites. Opposed to this suggestion is the fact that no large intrusives of peridotite reach the upper crust of the earth. If the earth is formed by the accumulation of such materials as meteorites, and if the accumulation was rapid enough to cause general fusion, much of the peridotite may have settled to form a deep shell. But the nature of the material from which the earth grew is uncertain, and the fusion of the whole surface is yet far from being proved or disproved. The deeper parts of the earth may be as heterogeneous as the rocks of the upper crust. The idea of a peridotite zone, then should be considered as one good working hypothesis in a group of several possibilities.

The probable thickness of such a deep stony shell is naturally more uncertain than its existence. Measurements of the earth show that it is a flattened spheroid, and a calculation (based on some further assumptions, however) suggests that the outer stony portion of the earth is roughly 1,000 miles thick.² Various other speculations and calculations give results of the same order of magnitude. Earthquake data indicate a change of material or condition at some such depth.

Below the rock shell just discussed and probably several hundred miles thick is the core or nucleus of the earth. This is almost certainly of different material from the lithosphere, and probably the most common suggestion is that it is metallic, as might be expected from the meteorites, from its magnetism and from its density.

If its segregation from the upper zone is not complete, there may be a gradation zone of stony iron (pallasite) of considerable thickness. The chief opposition to the idea of metallic material in the core comes from those who do not believe that the earth ever was fused to a sufficient degree to permit such segregation. There is the further fact that earthquake waves passing through the core not only behave differently from those in the higher zones but their record is *illegible*. The waves do not simply change to the speed expected in heavy material like iron, they become wholly anomalous. Possibly the core is a heterogeneous

¹ WAGNER, P. A., South African Jour. Sci., vol. 25, p. 127, 1928.

² WEICHERT, "Our Present Knowledge of the Earth," Smithsonian, 1909. ADAMS and WILLIAMSON, Distribution of iron in meteorites and in the earth, Washington Acad. Sci. Jour., vol. 14, p. 335, 1924.

mass that made the original nucleus of the earth; possibly the core is itself zoned; or possibly the iron in it is molten; or possibly pressure and heat cause its material to take on an "electronic" condition with a great change of properties—some form of matter not familiar to physicists except in the stars.

Quite aside from the study of the changes in the *material* in different zones is the conclusion that the *physical condition* of the earth must be different at different depths. The geodetic data led to the idea of an asthenosphere below the superficial zone of strong rocks. The lower limit of the asthenosphere is even less definite than the upper limit, but Barrell¹ suggests the following mode of attack:

For the study of the deeper zone the significant data are those of folding and thrust faulting. Mountain-making folds are not well explained by geodetic data and calculations; the orogenic movements seem to be related to deeper, more fundamental earth movements with only minor adjustments of load. If one does not accept the hypothesis of drifting continents, the simplest explanation of folds and thrusts is that some portion of the interior of the earth changed in volume. If the core shrank, the outer crust crumpled and shortened to fit the core. The periodicity of the diastrophism shows that the stresses accumulated gradually, and after a large accumulation the earth yielded during a period now recognized as a "revolution." After that the stresses again accumulated. (But see page 159.)

It is clear that the part of the earth that is condensing can not be the part that resists condensation and accumulates the stress. The condensation must occur in an inner core. The resistance would then be stored up as an elastic strain in a surrounding shell.²

How much is core? Barrell notes that if the core has a radius one-half that of the earth (one-eighth of the volume), then the whole earth, having its radius shortened 1 per cent since Cambrian time, must have decreased in volume 3 per cent, and the core, *where it was condensed, lost 24 per cent*—too much to seem reasonable. As a second assumption take the condensation as distributed over the inner three-fourths of the radius, and the inner portion will be condensed 7.1 per cent, a more reasonable estimate. A core with a radius three-fourths of the total radius is of the same order of magnitude as the core inside the supposed stony shells of earth (page 180), and the estimate has the merit of being based on wholly different data.

If the idea of a shrinking core in the earth is acceptable as a cause of orogeny, it would seem that a shell hundreds of miles thick and with a high degree of strength should exist to store up the stresses. Barrell suggested that such a shell be called

¹ BARRELL, JOSEPH, The spheres of earth and their effects on the lithosphere, *Am. Jour. Sci.*, 5th ser., vol. 10, pp. 499-529, 1925.

² CHAMBERLIN, T. C., Carnegie Inst. Washington Yearbook 25, 1925-1926, says that there has always been a really competent resistance deep within continents.

an "akthosphere" and estimated that it extends from the asthenosphere down to the "barysphere," perhaps 1,000 km. The barysphere constitutes the shrinking core. Willis prefers the terms stereosphere and centrosphere.

Is There a Magma Zone in the Earth?

Many geologists have speculated as to the probability of a magma zone (sima) somewhat widely spread around the earth below its crust (the sial of the continents).¹ The first suggestion comes from the wide distribution of magmas and from the extrapolation of the geothermal gradient; the gradient might easily cross the fusion curve so as to result in a fused zone. Confirmation is found in the study of heat supplies. Opposed to the idea is the known rigidity of the earth, but the advocates of a magma zone attribute the rigidity to pressure.

Joly and Holmes, in a study of radioactive heat (page 147) and the conductivity of rocks (page 151), concluded that the temperature must rise to the fusion point of the rocks a few kilometers in depth. Holmes estimated that it would take 30,000,000 to 40,000,000 years to melt the basalt zone under its probable condition in the earth.² Joly adds the important idea that the mechanism of cooling after fusion involves a drifting of continents, so that the heat of the magma zone may be absorbed by the oceans. This idea is relatively new and needs further testing, but it has not fulfilled the high hopes of its originators that it would be generally acceptable. It assumes a zone of basalt and one of peridotite, and further that the radioactivity at depth can be fairly estimated from the character of basalt and peridotite near the surface. These postulates are not universally acceptable, and it is still possible that at the depths under discussion some condition may reverse the reactions or absorb the heat. The idea of periodic melting should be retained as one of a group of hypotheses.

A second method by which the earth might have acquired a molten zone is the rapid infall of planetesimals in the course of its growth (page 146), but this rate may only be inferred from scanty remote evidences. The widespread dominance of igneous action in the Archean is probably the point most suggestive of a completely fused surface at some early stage. Other evidences are found in the widespread uniformity in chemical characters of igneous rocks, in the limited depth of density variations in the crust, in the small quantity of salt in the sea, and in the rotation periods of the moon and planets. The authors of the planetesimal hypothesis regard the evidence as insufficient and believe the growth of the earth was slow. Their evidence is no more conclusive than that of their opponents. One noteworthy argument against any general

¹ DALY, R. A., The outer shells of the earth, *Am. Jour. Sci.*, 5th ser., vol. 15, pp. 108-109, 1928.

² HOLMES, ARTHUR, Radioactivity and the earth's thermal history, *Geol. Mag.*, vol. 62, pp. 529-544, 1925.

molten condition is that dissolved gases would rapidly escape from such a magma, leaving the earth's rocks and later magmas almost free from gas, whereas we know that modern magmas are still highly charged with gases.

It may also be noted that any general fusion of the earth or its surface would be expected to mix the fluid so that broad petrographic provinces would not differ much. A study of the sun and of Jupiter, however, indicates a considerable permanence of heterogeneity even in a fluid rotating globe.

As already noted, those who maintain that there is a magma zone attribute the earth's rigidity to pressure on the liquid, for it is known that pressure greatly increases viscosity.¹ This may be more easily understood by analogy with glass, which in the parlance of physical chemists may be a supercooled liquid. To most men glass seems rigid and solid, and in some such way pressure might make a magma seem rigid and solid. Most of the tests applied to determine such a matter have been of short duration. No experiments seem to have determined the strength of a fluid with pressure-induced viscosity under moderate, long-continued stresses, but it is generally agreed that such a fluid would be very weak. Two experiments have some bearing. Adams² tested glass at 800°C. (one thousand million times as viscous as water) in which convection is *slight*, negligible, and finds it can be cooled 300° to 500°C. and at that temperature still yields to stresses of a few moments' duration. Other glasses might pass from the temperature of convection to rigidity in 200°C. but probably no less. On the assumption that the earth was once molten, a mass at a depth of 600 km. may have cooled since convection *ceased* nearly 200°, but below that there is little chance that it cooled 200°; hence its rigidity is probably due to crystallization, but it no doubt has reservoirs of liquid. The conclusion would be more certain if data were available on the strength of fluids under high pressures and subjected to long-continuous stresses. Another test of considerable significance is that of compressibility basalt glass,³ which indicates that the velocity of earthquake waves below the crust can not be explained by basalt glass, but requires a basic crystalline rock.⁴

Finally, there is the certainty of repeated base leveling (page 162). Such erosion must almost certainly produce some prolonged stresses in the rocks below the crust, and to most petrologists these evidences of strength

¹ The work of Bridgman at Harvard indicated high viscosities under pressure, but it is not certain that under such pressures the liquid remained free from crystals. See N. L. Bowen, "The Evolution of the Igneous Rocks," pp. 312-313.

² ADAMS, L. H., Temperatures at moderate depths within the earth, Washington Acad. Sci. Jour., vol. 14, pp. 464-467.

³ ADAMS, and GIBSON, Compressibilities of dunite and of basalt glass, Nat. Acad. Sci. Proc., vol. 12, p. 282, 1926.

⁴ DALY, R. A., Am. Jour. Sci., 5th ser., vol. 15, pp. 116-118, 1928, notes that Bridgman finds glass at low temperatures abnormal and conclusions from it uncertain.

and elasticity and rigidity indicate a crystalline solid earth, except for local magmatic reservoirs.

SUMMARY OF THE RELATION OF MAGMAS TO THE INTERIOR OF THE EARTH

The little that we know of the interior of the earth as a source of magmas is summarized above, and some detailed suggestions as to deeper conditions are extrapolated, but emphasis should be put on the fact that we do not know. The details of the several pictures have been differently combined by different petrologists. It is hard to be fair to the authors of a comprehensive hypothesis in a brief review of their ideas, but it is believed that certain summary statements may help the student to draw a contrast. The summaries are therefore given here without quoting or giving credit, but a series of references leads the student to the statements made by the advocates of the several ideas, where their own words state their position in each matter. Briefly magmas may be residual either from a molten earth, from an early molten surface or zone or from a periodically melted zone; or they may be generated either locally where the original planetesimal was most fusible, or in the supposed asthenosphere at times of orogeny and friction, or in the supposed basalt and peridotite zones by radioactive heat.

1. The earth may have been practically solid at all stages of its planetesimal growth and history, eliminating by igneous action any relatively small bodies of fluid formed in it. The "crust" is riveted by igneous rocks and is now firmer than ever before. Diastrophism is a periodic yielding down to a shrinking core, with a wedgelike development of mountain units, which may be several hundred miles deep. In this theory little attention is given to the differing densities of the rocks in different areas. Such broad differences as are found are attributed to differences in the planetesimals in different parts of the earth. The petrogenic problem is one of *ascent* or *synthesis* of rocks from various planetesimals. Some of the more fusible masses, deeply blanketed, would be the first to form magmas. Fundamentally *the place where a magma is formed is related to the place where the planetesimal fell and its nature. The variety of igneous rocks is a primary feature.*¹ Secondly the processes of orogeny and differentiation may modify the rocks. Magmas work their way upward because of their low density compared with their solid surroundings, but there is less tendency for heavy minerals to rise than for light ones, so that the long continuance of rising magmas may result in some residual concentration of heavy rocks in greater depths.

¹ CHAMBERLIN, T. C., and R. D. SALISBURY, "Textbook of Geology." T. C. CHAMBERLIN, Geol. Soc. America Bull., vol. 32, p. 208, 1921.

2. A second notion is that of a magma layer between a solid crust and a solid center, both solid parts probably consisting of several shells or zones more or less continuous. The crust practically floats on the magma zone. Some who uphold this idea consider the magma a residue from a wholly fused earth, others a residue from a superficial fused zone. Continental masses may drift upon this "sima" of magma below.¹ Mountain-making movements are results of continental drift against an obstacle.

Under this theory the petrogenic problem is one of *descent* since the great zone of liquid might have become thoroughly mixed and stabilized. It is not at all certain, however, that an earth showing density stratification now could ever have been stirred up to approximate uniformity of magma. *The variety in rocks* in a broad way would be a result of *differentiation* rather than of synthesis. Batholiths rise from the magma zone through fractures, developing plutonic wedges. The magma may be modified by assimilation of its wall rocks and roof.

The evidences of the solidity of the main part of the earth during the last few years, when studies have been made of earthquakes, tides and shifting of the earth's axis, are convincing most petrologists that this magma layer has not been present during this particular period if ever. The petrographic study of continents supposed to have drifted apart on the "sima" indicates that they are little related, and the form of the continental edges does not agree as well as might be thought from a glance at the map. A final argument against the idea is the evidence of the moon (page 208), which very probably had a magma layer near the surface at one time, and must have had as much reason as the earth to have masses drift around, yet shows no folded mountains.

A variant from this history is embodied in an old suggestion that the fused surface may have cooled so as to form crusts, which foundered and accumulated at depth to form a spongy solid mass, in which the magma remained as an interstitial fluid. Such a structure is very doubtful, but the petrogenic problem would be one of descent.

3. It is possible that a once liquid earth or an earth having a liquid surface layer became almost wholly solidified, differentiating after the manner of large known intrusive magmas, probably by settling of crystals, so that successive zones developed with heavy rocks at great depths. If the magma was ever continuous around the earth, it would produce fairly uniform shells, disturbed only by later igneous action and orogeny.

After the solidification of the crust in the manner just described, any generation of magma would be restricted to deeper and deeper zones as the surface cooled. The *petrogenic problem involves* (1) *the depth of magma generation*, the deeper zones being composed of more basic material, and (2) *a later descent by differentiation*. The variety of material resulting from depths of formation is probably not great com-

¹ DALY, R. A., "Our Mobile Earth."

pared to that which results from differentiation. The crust for a depth of 50 km. has probably, in general, temperatures considerably below fusion. At some depth below 50 km. in the supposed asthenosphere, the temperature is close to the temperature of fusion. The least accession of heat or relief of pressure will generate magma. No large reservoirs could be held by the asthenosphere, however. The small accumulations, lighter than their walls by reason of their fluid state, would rise, slowly at first but converging into larger streams with more rapid flow, as streams of water from glacial melting unite near the base of the glacier into rivers. Only as magma reached the lithosphere would it find a wall with strength to contain a large reservoir. The rigidity of the asthenosphere is maintained by continuous elimination of the fluids formed. The asthenosphere, at depths of possibly 60 to 600 km., is the special generator of magma.

As a variant from this history and others it may be noted that widely spreading batholiths invading a solid earth, such as are suggested by the authors of the planetesimal hypothesis, may overlap, so that after a long time the combined batholiths form a rough zone. The results might be similar, but the zones would be less regular.

4. The earth may be solid as now supposed but instead of having cooled from an early heated state may be periodically heated by atomic disintegration so as to develop molten zones. On the deeper molten zones the upper "crust" or continental masses may drift along so far as to allow the molten layer below to be cooled to solidification by ocean waters.¹ Then the heating begins again in a new cycle. Orogeny may be attributed to drift against an obstacle. This theory is consistent with the idea of shells of different material in the earth and with the expected gradation from one to another, but postulates a concentration of granite in and under the continents. This concentration of granite impresses minor irregularities on the zones of basalt and peridotite supposed to lie below (Fig. 128). This relatively new idea has not been developed to clear up its relation to isostasy.

The petrogenic problem involves chiefly the depth at which fusion occurs, whether in the zone of granite, basalt or peridotite, and so far as outlined the theory largely ignores the known differentiation in single masses of magma. A granite batholith is attributed to fusion of the deeper parts of the granite continent. There is no reference to the development of granite tops from the large masses of basalt fused at greater depths over wide areas. The basalt zone is the special generator of magma. Most of the granite is so close to the surface that its high radium content does not suffice to heat it faster than the heat is conducted to the surface and dissipated. On the other hand, the deeper peridotite zone, though better blanketed than the basalt, has less radium and takes several times as long to evolve heat enough to melt.

¹ JOLY, JOHN, "The Surface History of the Earth," Clarendon Press, 1925; R. A. DALY, "Our Mobile Earth," Charles Scribner's Sons, 1926.

IMPORTANT FACTORS IN MAGMA GENERATION

In review of the theories just outlined it is noteworthy that they assign the dominant role to different processes. (1) The heat of compression of accumulating planetesimals is diffused and general rather than local, and develops local magma only where there is locally some particularly fusible planetesimal material. (2) On the contrary, the heat due to orogenic movement is strongly localized, and may be great even at depths far down toward the centrosphere. Deep adjustments, caused by shifting surface loads, affect chiefly the lithosphere and asthenosphere, though these have ill-defined limits. The friction from such movements is not commonly concentrated in restricted regions but is widely spread. If magma forms a considerable accumulation, it is a result of the rise of innumerable small scattered masses to a reservoir near the base of the lithosphere. (3) Finally, the heat of atomic disintegration is widely diffused and although the heat is great in quantity, it should be expected to fuse broad regions of the earth's basalt zone. Localized igneous action may be attributed to drift of a continent against a barrier, involving a good deal of friction, or possibly to some process by which radioactive materials become localized.

The common association of batholiths with mountain folding and the sequence of invasion soon after orogeny (page 172) or alternating with it, are eloquent of the dominance of orogenic friction as a local source of heat. It is not suggested that this source alone serves as a generator of magmas, but added to the others it seems likely that orogeny "pulls the trigger" in many cases. One must here guard against confusing cause and effect. The absence of mountain folding in connection with other batholiths is equally eloquent testimony that the friction of orogeny is not absolutely vital to the production of magma. The heat of atomic disintegration, that of compression, and that of friction during rock flow may unite to produce the great batholiths under the plateau regions.

Quantitatively the heat of atomic disintegration seems to be ample without appeal to any of the other factors, and it is chiefly the sequence of batholiths after folding which indicates that radioactivity is not the dominant factor in the common orogenic batholith.

SUBCRUSTAL FUSION AND FUSION IN THE CRUST

The "crust of the earth" is an expression with very loose significance. Perhaps it is best used for the upper 10 miles of the lithosphere or for those portions of the earth which are so near the surface that erosion such as has occurred since early Archean time might expose them to view.

Many suggestions have been made from 1780 to the present time that masses of the crust have been fused almost in place. The idea has been less and less favored, however, because none of the rocks tested have a

composition favoring such an origin. It seems likely that a magma already fused may dissolve some rocks in the crust, but it is very doubtful whether heat generated in rocks or heat added from a magma or any other source will fuse any large mass of rock in the crust. Much is being written of granitization as a process that takes place *near* batholiths, but it is probably more largely injection and solution than fusion, (see pseudotachylites and palingenesis, pp. 116, 147, and 413). As noted on pages 217 to 221 and 421 to 424, the emanations from a magma may cause changes in the composition of the wall rocks. Where granite intrudes schists the rock is injected *lit-par-lit*, but in a wall rock of quartzite the changes are pervasive rather than in layers, and the altered rock is almost as massive as the granite. It is probable also that although the introduced material is commonly a dilute hydrous emanation from the magma, the emanation may be followed in some places by the main magma itself. If a granite magma is introduced into massive quartzites and so disseminated that its channels of access are not visible, it may so largely replace the original as to form a rock indistinguishable from granite. Some intermediate stages in this series of replacement effects suggest that certain large masses of granite may be really results of replacement.

It may be well before leaving the subject to be as specific as possible about the probable depth of magma generation and the depth of reservoirs. Under the planetesimal hypothesis, if no fused zone ever developed, magmas may have been formed at any depth below that at which the geothermal gradient approaches the fusion curve. Under the hypothesis of an asthenosphere, the magmas are generated at depths estimated to be more than 60 km. but probably not more than 600 km., and to accumulate in reservoirs at depths probably less than 50 km. Under the hypothesis of periodic fusion by atomic disintegration, the magma is generated in the basalt-eclogite zone at depths around 10 to 30 kilometers in most instances and much less often above or below that depth. All these depths are "subcrustal."

Most of this discussion as to the depth of magma is based on the conclusions derived from extrapolated curves, more or less confirmed by the data of geodesy and atomic disintegration. One further line of argument may be of interest.

An attempt to estimate the depth of conduits was made by a calculation based on Mauna Loa at 14,000 feet and Kilauea at 4,000 feet above sea level, the two being only 16 miles apart. Both of these craters are open, but they rarely erupt at the same time. They are probably not connected, but possibly one is fed from a satellitic reservoir; Iddings¹ reports, however, that in 1872 both were especially active at the same time. *If the difference in heights of craters is due to differences in vesiculation, the depths to which the columns are separated can be estimated.* With increasing vesiculation near the top a reasonable difference in gravity for the two columns would be 3.0 minus 2.7. The lighter column should be 1.1 times as high as the heavier. The

¹ IDDINGS, J. P., "Problems of Vulcanism," p. 18, Yale Univ. Press, 1914.

0.1 greater height is 10,000 feet. Below the top of Kilauea there should be 100,000 feet or nearly 20 miles of column to the joint reservoir.

On a small scale such a condition is illustrated by two fountains at Vesuvius, both open at 300 feet difference in level and the upper one the more active.

Still another evidence that magmas must form at depths of more than 10 miles is based on a calculation of the conduction of heat to the surface.

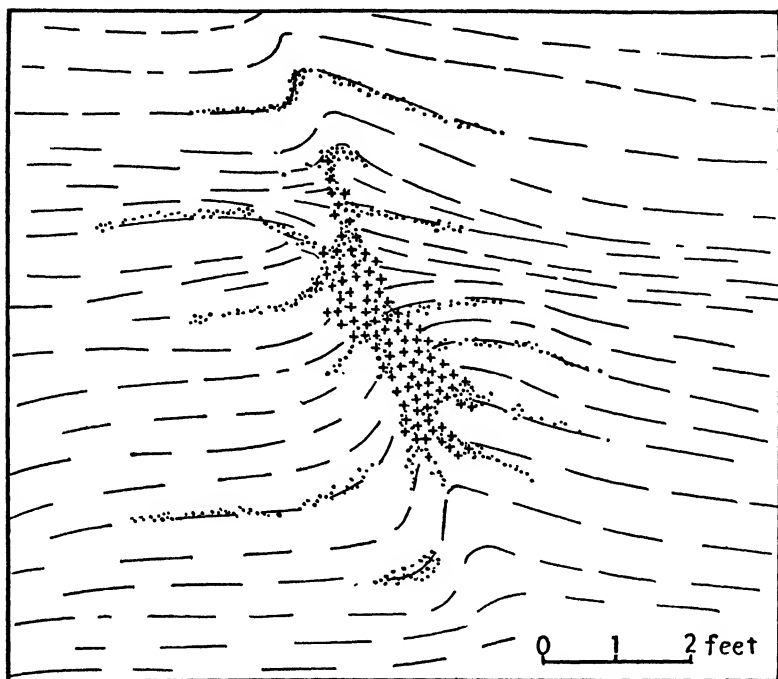


FIG. 129.—Sketch of sediments altered in the deep zone near Georgian Bay, Ontario. Where the fold is sharp and some beds seem to have been broken there is a pegmatite that may be a result of extreme metamorphism, a secondary granite or segregation granite generated almost in place.

Heat probably is rarely generated fast enough within 10 miles of the surface to produce magmatic temperatures in spite of that conduction.

Finally, the great erosion, to depths of more than 10 miles in many places, has exposed rocks that existed for ages at such depths (Fig. 247), and no fusion in place is indicated by these rocks.

Two features may be cited, however, as indicating that certain small amounts of magma may be generated in the crust. Local patches of granite in a metamorphic sediment, especially if the granite is pegmatitic, may be attributed to crystallization from solution in connate waters (Fig. 129), but such patches are rare except in contact zones or the roof rocks of batholiths. A second indication is found in certain conglomerates that seem to have flowed into dikes or apophyses. See Fig. 130 which shows the matrix of a conglomerate seeming to send apophyses

into a lamprophyr. This is in the contact zone of a stock. Thus the two cases that most strongly indicate fusion of the crust are so related to the heat of near-by igneous intrusives that there may be some suspicion that



FIG. 130.—Lamprophyr dike, Snowbank Lake, Minnesota. The dike crosses a conglomerate, and the matrix of the conglomerate injects fractures in the dike. Since both are in the contact zone of a syenite stock, the lamprophyr may have been injected when the conglomerate was almost melted.

fluxing emanations as well as heat may have been contributed to the magma formed. Most generation of magma probably occurred below the crust.

NATURE OF "PRIMARY MAGMA"

Many years ago, while studying the lavas of Iceland, Bunsen decided that there were two main magmas, granitic and basaltic, and that rocks of other composition were mixtures or secondary. As recently as 1911 published papers have maintained such a theory. Rosenbusch grouped rocks in three series and suggested differences between them based on primary differences in the magmas. A study of Cordilleran intrusives suggested that there were at least three different magma supplies—whether primary or subsidiary reservoirs. The frequency of occurrence of rock types (page 126) suggests that two magmas

at most will account for the known variations of igneous rock composition.

Meanwhile a number of men, possibly impressed by the calculations of average igneous rock, seem to have concluded that the primary magma of any common igneous series is a rock of monzonitic "intermediate" composition. From such a magma a series of differentiates might range from gabbro to granite. Von Richthofen in California emphasized the sequence of flows at a volcano: 1, intermediate; 2, more siliceous; 3, more basic; 4 and 5, *increasing divergence*. He naturally assumed that the primary magma was intermediate and gradually split into extremes; but the whole idea is open to question. Von Richthofen's sequence is not so prevalent as he supposed; in any large compilation the sequence is erratic (page 136).

Furthermore, there is no good reason to pick as intermediate rocks those between granite and basalt. The true extremes are silicite and ores. In such a series basalt is intermediate. In sharp contrast to the other ideas, then, there is the suggestion that the primary magma of most igneous rocks was *basaltic*.¹

1. Experimental work shows that basaltic magma will easily yield granitic residues, but no experiment indicates that basalt can be formed from granitic magma.

2. Large basalt intrusives have differentiated to granite tops; small ones have not. It seems likely, therefore, that the still larger batholiths would have such large granite tops that no other part would be seen, and so they commonly have.

3. In contrast, the great flows that seem to have come rapidly in large quantity from great depths are all basaltic or andesitic.

4. Every great igneous province has basalt, and each geologic period of igneous activity has basalt.

5. Intermediate rocks, like the "average rock," are relatively scarce.

The arguments seem conclusive, leaving unanswered only the question how the asthenosphere or region of magma generation supplies such a constant product.

What Magma Would the Deeper Parts of the Earth Supply?—In early pre-Cambrian time magma may have been widely formed at or near the surface, but since the time of the basaltic flows of the Archean probably most magma has formed at considerable depth. The supposed materials are listed on page 152.

Generation of magma occurs somewhere in these silicate shells or mixed planetesimal materials. If there are shells, they grade into each other, and the composition of the magma will vary with the *depth of formation*. If planetesimal material has not formed zones, the later-

¹ DALY, R. A., "Igneous Rocks and Their Origin," pp. 171 and 315, McGraw-Hill Book Company, Inc., 1914.

generated magma would vary not only with depth but with lateral variation of the material. Finally, it would vary because there is a partial rather than a complete fusion of the rocks; any fusible portion of the rocks is probably drained out to accumulate at the base of the lithosphere.

This last point may perhaps be more significant than all the others together. Which are the most common minerals of earth's deeper shells, and which mixtures fuse at especially low temperatures? The tables note several (page 152) that melt below 1000°C. The ferromagnesian olivine, pyroxene, amphibole, and magnetite melt around 1200°C., and feldspars from 1100° to 1500°C., but these temperatures are much reduced by mixtures and by additions of water (pages 154 and 236). It seems likely that the first portions of a rock to melt might be generally basaltic in nature, but there is little reason to expect them to be uniform. Any water available would enter the melt and facilitate melting. Fractional solution of this sort makes it likely that basaltic magma can be derived even from a zone with peridotitic composition.

The process of batholithic accumulation might influence the degree of uniformity. If there are many "roots," the scattered supplies might be expected to produce a mixture much like that in other regions; if the whole was derived from an especially fusible planetesimal mass, it might be notably different from any other.

How Uniform Is Primary Basalt?—The material best illustrative of primary magma is probably the plateau basalt, which seems to have come rapidly in large volumes from great depths. The flows range rather widely in a single field and in the averages of different fields. The alkali content ranges from 2 to 5 per cent, and the silica ranges so widely that there may be 10 per cent free quartz in some and 10 per cent olivine in others.

An earlier section notes that certain petrographic provinces of great size differ so markedly from other large provinces that the magma of the two regions must have differed originally. The differences between the rocks of continents and those of ocean basins also suggest differences in the original magma. If we accept the geodetic data, the continents and ocean basins are not far from being in equilibrium, and the observed differences in density must extend to great depths to produce such a compensation, so great that one could hardly be an upper differentiate from a magma that yielded the other—depths of the order of magnitude of 100 kilometers or more.

Probably most large bodies of newly generated magma more or less closely approach basalt in composition, but different supplies vary at least as much as the variation in the andesite-basalt series. (See the list of further readings.)

MOTION OF MAGMAS

CAUSES OF MOTION; FORCES INVOLVED

Magmas move because of direct gravitative pressure, because of indirect gravitative thrust probably due to shrinkage of the earth's core, because of local changes in temperature and because of changes of state. Additional causes suggested include tidal strains; percolating water, heated and expanding; capillarity; assimilation and the principle of dilatancy. See the list of further readings.

1. Direct Gravity Effects.—Pressure on a liquid at rest is hydrostatic, that is, it is transmitted in all directions equally; but in columns of liquid several miles high the bottom pressure is, of course, not transmitted to the top. If a column of magma extends up a fissure from a magma chamber toward the surface, the height to which it will rise depends largely on the weight of the column and the weight of the column of rock resting on the reservoir. Normally solid rock is believed to be heavier than liquid of the same composition, so in case the earth's crust is basalt, a basalt lava would tend to be extruded.

The higher the columns involved the greater the tendency of the lighter one to rise, but even a short column would rise in the asthenosphere, moving slowly until the column attained considerable height.

ROCK AND MAGMA DENSITIES, ESTIMATED

Materials	Solid			Liquid	
	0°	1000°	1100°	1000°	1100°
Average sediments.	2.5 ±	2 42	2 41		
Average granite	2 75±	2 63	2 62	2 40	2 39
Average gabbro	3.00±	2 92	2 91	2 75	2 74

The change from solid to liquid is believed to reduce the specific gravity something like 9 per cent for silicic rocks and 6 per cent for basic rocks, but authorities disagree. The changes from temperature alone are relatively small, but those from crystallization not any too certain.

It seems likely that practically any igneous magma would rise in a fissure in igneous rock but might not rise through lighter sediments. Also that nearly any rock included in magma would sink from the roof, but some might rise from the floor. This is the basis of the idea of stopping and underhand stopping.

Besides the gravity differences between magma and solid rock there are specific-gravity differences that may develop in the magma itself by

temperature changes, vesiculation, crystallization and solution of materials. These may cause internal circulatory movements (page 206). Certain movements at great depths may result from the compression or expansion of magmas as loads are applied or removed.

2. Lateral Thrusts.—The shrinkage of the earth's core is supposed to be the cause of periodical readjustments by crumpling of the lithosphere. Great lateral thrusts develop at the borders of diastrophic units and are very likely to move magmas far and wide. These forces are possibly as great as the direct weight of rock on a reservoir. If a load of rock or magma is thrust into a region, an undertow in the asthenosphere may involve some further movement of magma. Crust movements and magma movements are coordinate, but neither can yet be shown to be the cause of the other.

The suggestion of a relation between the kind of magma and the kind of earth movements has already been referred to (pages 130 and 131).

A final suggestion as to lateral thrusts and magma movements involves the drifting of continents on a magma layer¹ (page 182).

3. Changes of Temperature.—The expansion of matter on heating probably has little effect in moving magmas. In contrast with the notion of shrinking of the earth's core, one may imagine an expanding core causing a magmatic thrust upward. The main expansion related to magma comes at the time of the formation of magma at depth. There may, however, be a sudden expansion of the immediate roof of an intruded magma, so that shattered blocks are stoped in. The opposite effect, contraction on cooling, may change the volume of a newly crystallized rock some three per cent, producing cracks, which permit pegmatite dikes and other injections, thus indirectly affecting magma motion. The surface crusts of many sills and dikes have a network of cracks filled with coarser rocks from the center of the same mass.

4. Changes of State.—The gases held in a magma under pressure in a volcanic conduit can hardly be said to cause much movement. When the pressure becomes great enough to break the crust on a crater lake, or when pressure is relieved in any other way, allowing lava to move upward, its movement is greatly accelerated and modified by gases. The release of pressure causes the gases to separate from solution, and the high temperature causes a great increase in their volume. As they expand they develop large volumes of vesicular lava, and their sudden upward movement causes the most violent explosions known on earth. This movement of the magma by the expansive force of the gas can not be attributed to an increase of heat, for the gases probably cool during expansion and are no hotter than they were in the magma solution; the motion is due to a change of state.

¹ A convection in the magma substratum is suggested by A. Holmes, *Radioactivity and earth movements*, Geol. Soc. Glasgow Trans., vol. 18, p. 575, 1928-1929.

Day¹ suggests that all strictly volcanic activity, as distinct from deeper igneous action, is due to gases. If there is only steam, there are explosions as at Lassen Peak. If there are other gases, their reaction may develop heat enough to melt rock in a crater as at Kilauea. This is a quite different notion of volcanoes from the older idea of a central pipe from a deep source (page 205).

The dominance of gas is indicated by observations at Vesuvius,² where an approximate calculation shows seven times as much volume of water (as liquid) as of solid, ejected at one eruption.

5. Miscellaneous Forces.—Rows of volcanoes along a coast line early suggested that sea water might percolate into magma and expand to cause motion. The idea is not favored (page 212), though seepage may cause explosions, as has probably happened at Kilauea.

By assimilation of its walls a magma chamber may be enlarged and the magma thus move outward. There is also the process of gas fluxing as a mode of motion or penetration. The "streaming" gases (page 218) may be so concentrated locally as to be active in boring their way through other rocks.

It has been suggested that magma is pressed out of the centrosphere by tidal kneading. Darwin early calculated the value of tidal stress as 16 grams per square centimeter at the pole and 128 grams per square centimeter at the center of the earth, with intermediate values elsewhere. The *difference* in strains is 112 grams per square centimeter in 4,000 miles. The gradient is so low that little motion is to be expected.

Lit-par-lit injection should be compared to water soaking up by capillarity between sheets of paper and later freezing. This may be contact action, but the distinction of regional and contact metamorphism was made before it was recognized that nearly all the Archean gneisses overlie batholiths. These emanations favor schists, and the injection may penetrate miles laterally from outcrops of the main batholiths.

Mead has noted that movements along curving fractures may result in openings that draw fluids in from the adjoining wall rocks³—a process of dilatancy. Pegmatites may illustrate the action.

MODIFYING FACTORS

Viscosity.—The direct transmission of pressure throughout a liquid, noted in an earlier paragraph, is modified in moving magmas by viscosity. Magmas may be very highly viscous, especially if they are siliceous and not hydrous. The lavas of Hawaii are estimated to be twenty times as viscous as water. Rhyolites may flow when one thousand times as viscous as water.

¹ DAY, A. L., Some causes of volcanic activity, Franklin Inst. Centenary, p. 19, 1924.

² PERRET, Vesuvius eruption of 1906, Carnegie Inst. Washington Pub. 339, 1924.

³ MEAD, W. J., The principle of dilatancy, Jour. Geology, vol. 33, p. 697, 1925.

The viscosity probably determines whether or not an intrusive sheet will thicken into laccolithic form. A particularly viscous laccolithic magma may even fault up its roof, forming a bysmalith. An extreme example of magma movement when viscosity was enormous was afforded when the spine of Mount Pelee moved slowly up for a long time. The length of lava flows is inversely related to their viscosity; rhyolites do not spread in such lava fields as basalts.

The effects of viscosity in magmas are thus variously described in a series of papers listed in the further readings on Viscosity of magmas (page 488).

Harker thought diffusion had important effects in magmas but that magmas are too viscous to allow much crystal settling. Daly approved the idea of crystal settling in magmas but indicated that the magma is too viscous for diffusion or much deep-seated convection. Pirsson thought convection occurred in a laccolith but considered diffusion unimportant in such a viscous liquid. Thus we get around a circle, and all the processes mentioned deserve consideration quantitatively. Fortunately Bowen has supplied some laboratory data, and others have supplied a sufficiently close check by field observations, so that we now feel confident that diffusion is very slow (page 233), and that crystal settling and convection are still worthy of consideration.

Viscosity may also explain the relative scarcity of rhyolite flows, for the viscosity of siliceous magmas becomes high as it rises into cold rocks and loses its mineralizers. Basalt magmas would not be so affected by the change of conditions and might rise to form surface flows.

We must here distinguish viscous liquids and weak solids. In a pitch, a million of million times as viscous as water, stones will sink and corks rise in a few weeks. In a weak solid such as gelatin, gas bubbles and small stones will remain stationary indefinitely. All magmas that are known or have been approximated experimentally are liquids rather than gelatinous solids. Viscosity delays motion in the magma but does not stop it until viscosity rises and becomes almost infinite, for example, in the glasses.

Deep in the earth the viscosity of magmas is largely unknown. Most petrographers believe that magmas have only moderate viscosity, because narrow dikes have cut cold rocks and penetrated far into them; because water would be held in solution under pressure and it reduces viscosity;¹ because feldspars commonly show alternating zones, as if the magma had been a somewhat turbulent liquid, and because *lit-par-lit* injections have not crumpled the thin leaves of schist. On the other hand, there are experiments showing that pressure no greater than 10,000 atmospheres increases the viscosity of some liquids very notably, and certain field structures suggest that granites in particular are highly viscous before the final cessation of their intrusive motion. They can crack and allow aplites and lamprophyre to fill the cracks (Fig. 133); they can drag along their walls to produce the same orientation in the wall and the igneous rock; they support fragments and dismembered lamprophyres and schlieren as if in a highly viscous matrix; and the schlieren from heavy fragments do not run upward as if the fragment had settled. (See the list of further readings.)

Crystallization.—*Partial crystallization* may produce in a magma a sort of "mushy" consistency as a modification of the viscous state. It is

¹ George W. Morey (personal communication) thus estimates the effect: The viscosity at 500°C. of anhydrous potassium silicate, $K_2O \cdot 2SiO_2$ to $K_2O \cdot 4SiO_2$, would range from 10^{12} to 10^{15} poises; containing 10 to 15 per cent of water at that temperature it could not have had a viscosity above 1,000 poises and probably not more than 100.

estimated that a magma may still be eruptible with 50 per cent of crystals. In any event, density will be increased by crystallization, and internal movements may be initiated by this change in density (Fig. 131). The crystals themselves tend to settle or rise, depending on their density as compared with that of the magma.

The movements of a mass 80 per cent crystalline would probably be quite different. See piezocrystallization (page 59). Any general deformation of a region that included a magma at this stage of crystallization might squeeze out a residual liquid as from a filter press.

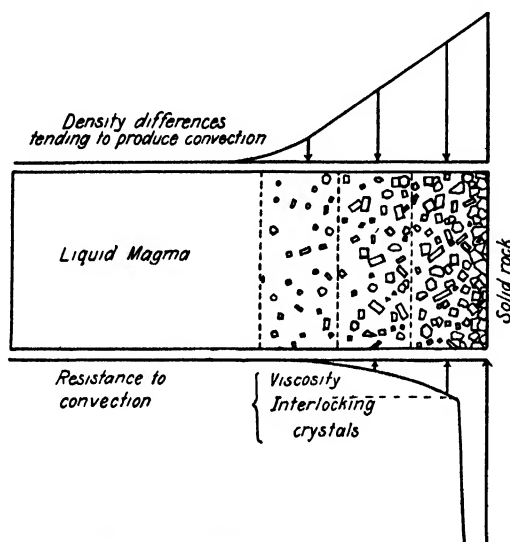


FIG. 131.—Diagrammatic section across the outer part of a magma chamber where the magma is crystallizing along a solid wall. Arrows suggest the probable forces producing and resisting convection.

A minor effect on motion may be attributed to the growth of a crystal from a saturated solution. It may be that some veins are widened by crystal growth. The leaves of *lit-par-lit* injections may be thus opened.

The Structures Resulting from Movements in Crystallizing Magmas.

It is after a magma becomes largely crystallized that structures of great interest have developed in some of the deeper intrusive granites studied by Cloos and by Balk (see pages 29 and 30). Nearly all granites show plane and linear structures if examined in detail, in spite of the common report of their massive character. So many proofs have been discovered of motion at this late stage in crystallization, and so many of the small masses studied moved at this late stage in the same general direction as at the early intrusive stage, that the European workers have come to believe that many granite magmas when originally intruded were more mushy than liquid. The early control, however, of the forms of some large masses may be seen by exposures of roof, walls or floor and by the large inclusions

in them. These features do not necessarily coincide with such structures as the orientation of minerals at a late stage of crystallization. A magma that stopped its way upward into some mountain folds may at a late stage have its crystals oriented by a recurrence of the lateral thrust that produced the folds, quite independently of the forces that guided the upward stopping. It is necessary, then, to be on guard against interpreting a late structure as an early control. If this precaution is observed and if the masses have small sizes and simple structures, the position of



FIG. 132.—Horizontal aplite dikes in a schist with vertical linear schistosity in the wall rock of a stock at Snowbank Lake, Minnesota. These are interpreted to mean that the viscous intrusive and this wall rock yielded together and that the yielding to the magmatic thrust was in a nearly vertical direction. Aplites then filled fissures about normal to the direction of least pressure—almost tension cracks.

crystals oriented by the motion of the mushy mass may give important data as to the source of an intrusion and the form it assumed. Some of the general results of Cloos's detailed structural studies are noteworthy.

1. When the magma has reached the "mushy" stage, the magmatic heat and emanations have been acting on the wall rock so long that *the magma and wall respond to stresses together*. Both are in such a condition that they may fracture or may glide along certain planes, but still have enough fluidity or enough interstitial liquid to fill the fractures or inject the glide planes.

2. *Flow lines*, thus coinciding in the igneous rock and its wall, indicate the direction of yielding under magmatic pressures. In studies thus far made on the intrusives of the deep zone certain masses show peculiar uniformities of orientation of flow lines. In some granites the

flow lines have a general constancy of strike but differ in the pitch. In other masses the pitch of the flow lines is almost uniform, but their directions differ. No very good explanation has been offered for such uniformities, but the field facts are of interest.

3. Although the direction of yielding is shown by the flow lines, there is commonly a system of open joints and aplite and lamprophyre dikes and ore veins in a plane at right angles to these lines. In some districts these may be more easily seen and followed than the flow lines. If they lie in the proper relation to the flow lines, they show a tension or a direction of least pressure, hence a direction of motion at right angles to the joints (Fig. 132).

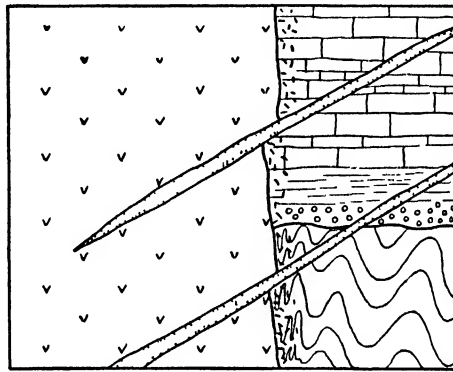


FIG. 133.—Aplite dikes (stippled) passing from the parent igneous mass into several wall rocks without much change in character. Note that the offset and the direction of motion to open the aplite fractures both suggest a tension or drag as the magma moved viscously upward.

4. Other joint systems and dikes may follow the flow layers in the rock, and in intrusives with nearly vertical walls there may be diagonal joints and aplites dipping from the wall into the intrusive. These are taken to be tension joints made while a viscous or mushy magma continued to drag along its walls (Fig. 133).

HOW MAGMAS MOVE

Motion below the Lithosphere.—The thickness of the lithosphere and the nature of the zones below it are highly speculative, and the forms of magma bodies have been variously conceived. Although it is possible that a basalt-magma zone exists all around the earth, it is hardly probable and if we ignore this idea, the magma must be considered as associated with crystalline rocks and tending to rise because liquids are lighter than similar materials solid at the same temperature. If the solid is weak, as in the supposed asthenosphere, rock flowage in the walls will permit the rise, but such motion will naturally be very slow until the volume of magma becomes very great. It seems likely that

small masses rise slowly through such weak zones and merge into considerable reservoirs near the base of the stronger lithosphere above.

As the transition from the asthenosphere to the lithosphere must be gradual, and as the conditions that determine the existence of an asthenosphere are similar to those in the roofs of batholithic magmas, it is possible that magma may rise slowly, by rock flowage in the roof and walls, high into the earth's crust. Nevertheless conditions must gradually change when the magma reaches the zone of abundant rock fractures. The probable features of the deeper zone were outlined on pages 171 and 172.

This rise of magma in the deeper parts of the earth may be modified by any other stresses, orogenic or isostatic, which move magmas. The common occurrence of batholiths in orogenic belts and their emplacement soon after folding suggest the common influence of thrusts. Some batholiths, however, occur in less-folded plateau regions, hence the orogenic movement is not essential to the rise of magma. The mechanics of the deep movement is probably less important than the fact that we have clear evidence that great masses of magma do rise into the lithosphere, apparently deriving their substance from deep roots. It is only since about 1890 that it has been generally agreed (and it is questioned even now) that granite batholiths have risen as wide masses from abyssal depths. They seem to have broken through and engulfed the foundations of the lithosphere, and locally and periodically they have reached levels in the crust that could later be exposed by erosion.

The relation of batholiths to mountain folds and thrusts suggests also that there may have been some nearly vertical fissure in which the batholithic magma, exerting a hydrostatic pressure, thrust the walls apart and so crumpled them as to give the impression of a great crustal shortening (page 159). A different impression, however, is obtained from a study of the internal structure of a number of batholiths, which indicates that the magma when partly solidified was itself subjected to the mountain-making thrusts. There was an alternation of injection and mashing, such that only the latest phases of intrusion escaped deformations (pages 439 and 440). Thus batholiths can be distinguished as "synchronous," if they accompanied folding, or "subsequent," if they followed the period of folding.¹

Intrusion of Batholiths in the Lithosphere.—The rise of some batholiths to points near the surface, by crowding apart the adjacent walls under conditions of rock flowage, is indicated in many places by a schistosity in the walls that boxes the compass around the edge of the mass (Fig. 134). These batholiths probably give us our best evidence as to the mechanics of magma rise at greater depths. The results are illus-

¹ BILLINGS, M., *Petrology of the North Conway quadrangle*, Am. Acad. Arts Sci. Proc., vol. 63, p. 134, 1928.

trated by the Black Hills of Dakota and by many stocks in the Archean shield areas.

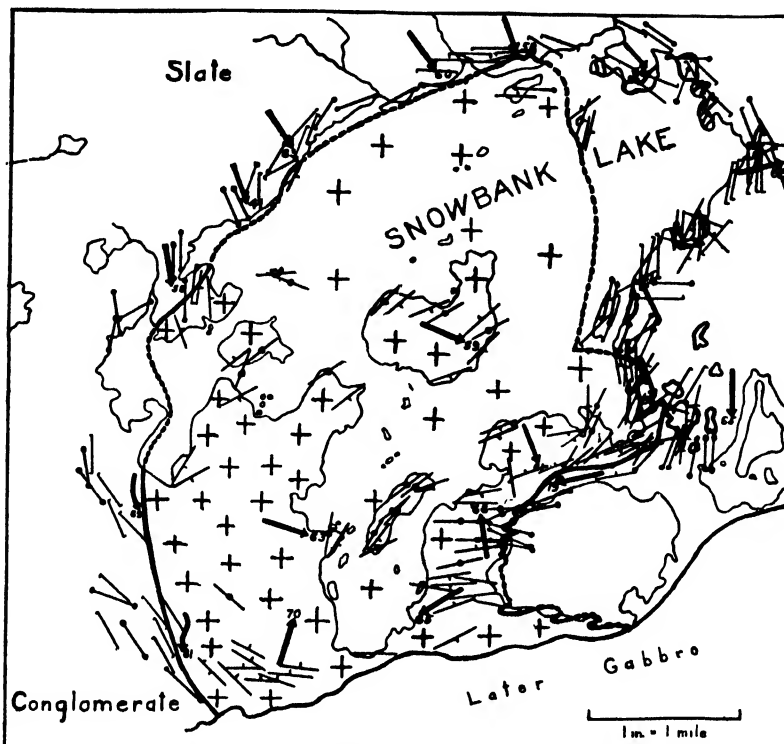


FIG. 134.—Snowbank Stock, Minnesota. The arrows show the projection of the structure lines both inside and outside the intrusive. Figures at the heads of the arrows show the pitch of these lines. The plane structures "box the compass" around the mass. Evidently the mass rose as an elliptical chimney leaning a little to the north as it rose.

In contrast to these the batholiths and other intrusives in plateau regions tend to have "blocky" contacts, which irregularly transgress the structure of the roofs (see Figs. 7, 135, and 144). They must rise at

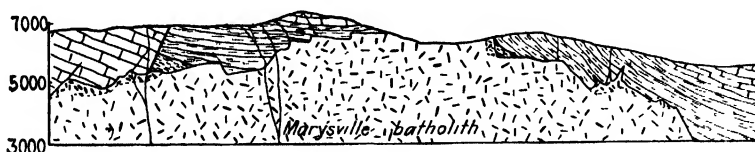


FIG. 135.—Cross section of the Marysville, Montana, batholith, about four and one half miles long. (After Barrell.) The blocky nature of the contact is noteworthy.

shallow depths by some quite different mechanism. The blocky nature of the contact is commonly found to depend on the nature of the wall rock and to be modified to approximate conformity with the structure in

particular formations (Fig. 137). It should be noted also that in a large majority of the great granite areas, the exploration of the contact as far as it has gone indicates an increased width at greater depths (see Figs. 126 and 127). There are many masses in the shield areas with nearly vertical walls, even where the geologic evidence indicates very great depths of erosion; and comparatively few granite masses of large area are known to have floors. The roofs are not domed or faulted enough to furnish good evidence of laccoliths or bysmaliths, nor have most batholiths a thick shale roof like laccoliths. As outlined on page 177, it seems likely that batholiths are connected by open magma with the

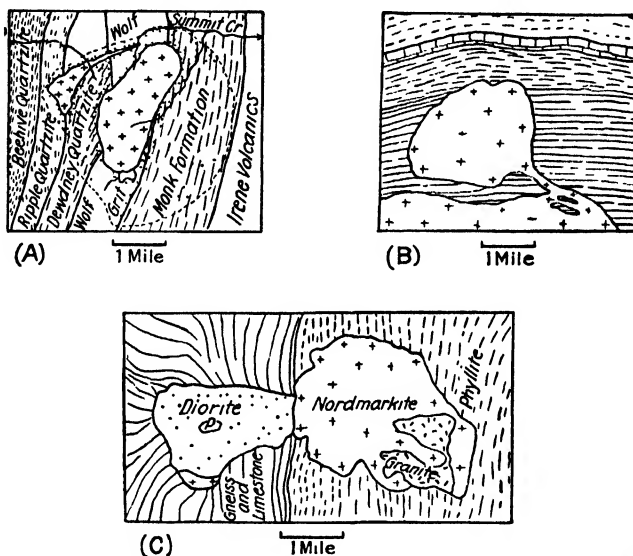


FIG. 136.—Stocks showing replacement of invaded formations by magmas. (A) Selkirk Mountains. (After Daly.) (B) Pyrenees Mountains. (After Bresson.) (C) Mount Ascutney. (After Daly.)

deeper chambers of magmatic accumulation. In many batholithic areas a great mass of material is missing (Fig. 136) and the walls are not much deformed nor the magma contaminated by assimilation. It appears that the missing rocks have dropped into the magma. This is the process of stopping.

Stopping is probably accomplished largely by the development of fractures in the roof and walls, caused by the sudden change of temperature when a magma moved into the neighborhood. Ring dikes and the structures of some intrusives suggest a cone of rock settling.¹ No batholith is known to have stoped its way to the actual surface, though it may be suspected that the roof foundered over some.

¹ RICHEY, J. E., Structural relations of the Mourne granites, *Geol. Soc. London Quart. Jour.*, vol. 83, pp. 653-688, 1927.

The results of stopping inside the magma are variously postulated. The stopped blocks may dissolve before moving far (and later differentiate), or may sink much farther and remain or dissolve at a depth so great as to remove them forever from the zone of observation. The abundance of inclusions near the floors of the large intrusives suggests that solution is too slow to make much difference in the upper crust.

Finally, it may be noted that some rocks are stopped and assimilated more readily than others (Fig. 137). Some well-cemented rocks may be stopped into a magma with consequent rise of the magma surface, but where the magma encounters loose, friable sandstone formations, it does

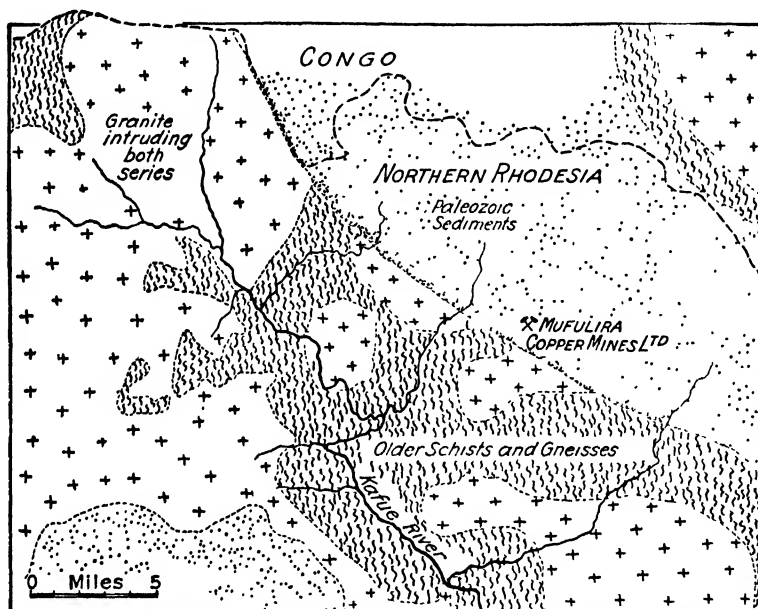


FIG. 137.—Map of Mufulira district, Northern Rhodesia. (After F. A. Gray.) Three granite stocks near Mufulira Mines intrude the Paleozoic sediments but invade them only a short distance, not forming blocky or rounded contacts as in the older schists.

not stop large blocks but stops and assimilates the sand grains and rises much more slowly if at all. For example, certain granites, intrusive into schists and sandstones in Northern Rhodesia, seem to have stopped abruptly within a few hundred feet after entering the copper-bearing sandstone.

Stocks apparently rise in the same way as small batholiths and may be equivalent to cupolas rising above the main curve of a batholithic roof. These projections of extra height may indicate a local intensity of stopping, or fluxing, or a local difference in the roof material.

The structure of the country rock may also determine the occurrence of other minor intrusions near batholiths without in any way modifying

our conclusions as to the rise of the main mass. It is recognized that granite masses may occur in sheets with a floor, but most of them can be traced to batholiths.

Smaller Intrusives.—The smaller intrusives seen so widely in the lithosphere—dikes, sills and laccoliths for the most part—are those which were probably injected by hydrostatic forces and followed the structure already present in the country rock. It is possible to distinguish theoretically those magmas that are “invasive” or “aggressive” in opening their walls from those “suctive” or “permissive” magmas that are drawn into openings created by orogenic movements, but the field criteria for the two are obscure.

Almost without exception magmas have followed planes of structure in the earlier rock. Bedding planes are responsible for sheets and laccoliths and in a small way probably for ribbon injections. Joints and faults are opened by intruding magma, and dikes and fissure eruptions fill the openings. Schistosity in a wall rock commonly guides a *lit-par-lit* injection. Intersections of nearly vertical joints, faults or other fissures may determine the position of pipes, chimneys, plugs or necks of igneous rock.

The kind of rock adjacent to the magma, quite aside from its specific gravity and structural openings, affects the motion of magma upwards. Laccoliths dome up their roofs almost wholly in thick shales, in some places with a more competent bed above the shale; and sills develop mostly in thin-bedded formations. Strong rocks may support an intrusion that domes its roof, whereas a relatively weak, thin crust may sag down under a load of intruded magma forming a lopolith (Fig. 14).

Pegmatites characteristically cross granites in fissures that have very indefinite boundaries, and it seems likely that at the time of their formation the wall rock had not yet completely solidified.

Extrusives.—It is also possible to distinguish, from the common structural control, some cases in which magma high in the lithosphere may have fluxed its way forward independently of the structure. For example, whereas it is known that some volcanic activity at the surface is a result of magma rising along a fissure, other volcanic eruptions result from a magma fluxing its way upward at least through its old crater crust if not from great depths. Still other periodic volcanoes, of course, open their craters by a different process outlined in the next paragraph.

Volcanic explosions are attributed to the presence of much water in a magma at a time when the pressure of the hydrostatic column or of the water vapor itself (page 213) has increased to a point where it forces the magma to break through its crust. The break releases the pressure, and water in solution rapidly separates as vapor and violently expands. The larger explosions enlarge the crater depressions into great *calderas*. If the explosion bursts upward from the crater, ash falls on the surrounding

area. If it emerges on the side of a cone, it may form hot blasts or ash flows, "*nuées ardentes*."

Quiet lava flows, on the contrary, seem to come to the surface mostly on account of the pressure of heavy rocks on the liquid below.

The structural control of magma movements below a volcano is uncertain. It has commonly been assumed that pipes or chimneys, in some places arranged along fissures, connect the crater lakes with a deep reservoir. The volcano Kilauea, however, has seemed to be typical, and careful observations make that structure under it improbable. In 1924 the lake was drained and blown out so that its deeper parts could be observed.¹ *It has no deep central pipe.* The bottom and most of the sides of the lake were found to be solid and cooler than the lava, and there were at least *three feeders of lava on the sides of the basin*, not in the bottom. The deeper connections of these feeders are of course unknown, but the observations raise a question whether or not the outcrops commonly called volcanic necks or plugs ever had any connection with a volcanic crater. Many such necks have a mixture of lava and breccia that certainly suggests a connection with surface explosion.

Large flows spread with an inclination of two or three degrees (Fig. 9). Volcanic conical mountains are built up more largely of fragmental material with smaller flows. Amygdaloidal and cellular structures are usually unrelated to the surface over which the flow moves, but if the surface is wet, some steam bubbles rise through the lava from the moisture.

Flows that reach the sea or a wet place have been known to develop ellipsoidal or pillow structures; the exact mechanism of the development is in some doubt. Something breaks up the moving flow into fragments still viscous enough so that they later get closely packed and almost reunited.

Ellipsoids furnish an excellent example of the process of development of a geologic theory. At first their origin was a geologic puzzle. Then a man saw a flow reach the sea and found that the cooled lava was ellipsoidal. For many years thereafter it was an accepted theory that ellipsoidal lavas were submarine flows. Then another man saw a flow on dry land become ellipsoidal. The criterion was ruined and the theory exploded. Since then wide and long studies have indicated that probably *most* ellipsoids form by the action of water on flows, but as a criterion of origin, ellipsoidal structure is simply suggestive, not at all conclusive. (See further readings.)

Possibly the explosive action of volcanoes, also sand flows and mud flows, may be logically considered as related to the nature of the surface of the country. It is suggested that tuff beds are made by incandescent sands in a flow in which semisolid grains are lubricated by the gases *being evolved*.²

¹ DAY, A. L., Some causes of volcanic activity, Franklin Inst. Centenary, p. 19, 1924.

² FENNER, C. N., Origin and mode of emplacement of the great tuff deposit of the Valley of Ten Thousand Smokes, Nat. Geog. Soc. Tech. Paper, Katmai 1, 1923.

Internal Movements in Magmas.—The fluxion and banded or layered structures in fresh igneous intrusives are hardly to be explained except by a motion of either intrusion or circulation. There are also many crystals zoned with alternating layers of different materials indicating a high degree of turbulence in some magma chambers. There are several possible causes of such turbulence. Stopped blocks may stir the magma as they settle and as the water boils out of them. Late injections may stir up the contents of a chamber already partly filled. Settling crystals and rising crystals not only move but drag along part of the magma as they move. Rising bubbles do the same. These movements are "two-phase convection"¹ (page 197). Thermal convection adds to the general effect.

Rates of Motion.—Some dikes that are very narrow, and have produced no notable contact effect, must have been injected with almost explosive suddenness to move thousands of feet before being chilled and solidified.² At the other extreme of magmatic motion was the spine of Mount Pélee, which rose slowly for several months. The rise of a stopping batholith may be extremely slow and still accomplish a good deal, because of the long time it takes such a mass of magma to cool. Convection circulation is observed in lava lakes at 2 to 5 km. an hour. A similar rate may be assumed in some deep-seated chambers. Lava flows and the blasts of ash from explosive volcanoes may move several miles an hour. It is recorded that at Hawaii a pahoehoe flow moved 11 miles an hour.³

Periodicity of Magma Action.—Lava flows are poured out intermittently. Intrusive action is probably similarly occasional. The reason for this intermittence is probably to be found in the rigidity of rocks and the gradual accumulation of stresses. One or more of several forces, such as isostatic stress, the deeper gravitative contraction of the earth's core and the pressure of accumulating gases, may grow stronger gradually until they exceed the strength of the rocks and an adjustment occurs catastrophically—in magmas by an eruption. At Kilauea the high levels of lava are reached at periods of 11.1 years, exactly the period of the sun spots.

Some explosions may be a result of water seepage, but it seems likely that the supply of gas for most explosive eruptions comes from the deeper parts of the magma.

Radioactivity heats the basaltic portions of the earth's deeper crust and may do so periodically; when the material is hot enough, the crisis comes in a melting and possibly a continental drift and oceanic cooling

¹ DALY, R. A., The nature of volcanic action, *Am. Assoc. Arts Sci. Proc.*, vol. 47, p. 76. F. F. GROUT, Two-phase convection in igneous magmas, *Jour. Geol.*, vol. 26, pp. 481-499, 1918.

² BARRELL, JOSEPH, The geology of Marysville, Montana, U. S. Geol. Survey, Prof. Paper 57, p. 158.

³ PALMER, H. S. quotes JAGGAR, Hawaii Volc. Observatory, *Bimonthly Bull.* 1, vol. 15, January, 1927.

(see page 182). Lava fields would be formed when the melting crisis comes. It is questionable whether such melting is world wide at any time.

Subsidence.—The periodicity involves not only an interval of quiet between eruptions, but in many cases a withdrawal of magma between eruptions. The craters of volcanoes overflow at times, but between times the level of a lava lake may be far below the rim of the crater. Many volcanoes after a long period of activity have so far subsided that great down-faulted calderas have been formed. A similar subsidence effect on a large scale is noted in large floored intrusions. The Sudbury, Bushveld, and Duluth intrusives lie in basins formed during or soon after their intrusion. The explanation is somewhat uncertain, but possibly the adjustments of stress may acquire some momentum and oscillate like a pendulum. The overflow of lava may be more than is needed to relieve the stress that started it. The deeper magma will then draw back after eruption. Changes of density due to cooling and crystallization may also cause changes of surface level, though these do not add any mass in a particular segment of the earth.

CYCLES OF IGNEOUS ACTION

The geologic records in a number of regions show a sequence of (1) flows, (2) intrusions of large size, (3) dikes. The sequence seems to be related to orogeny. The flows are characteristically basaltic and occur in regions where geosynclines have developed, erupting at about the time of beginning of an orogenic revolution. The sequence of flows is not very regular (pages 136 and 191) but commonly basalt forms the latest flow. During the folding igneous action seems less prominent, and the large bodies of magma in the deeper chambers have time enough to differentiate. After the folding the granitic magmas are intruded but not always in just the same regions as the flows. The batholithic cycle is probably not very clearly recorded, for the advance of the batholith obliterates the early small effects. The roof rock is no doubt first affected by hydrothermal emanation; aplite dikes and small intrusions follow, with *lit-par-lit* injections in schists; then comes the main advance of the magma, and only the late emanations and structures remain as a permanent record in the wall rocks. The solidification of the intrusives seems to make the crust more rigid, and the orogeny ends. Magmatic activity often closes with a renewed output of basalt, this time in dikes. In many places, after the granite intrusion but before the basic dikes are formed, there are intrusions of silicic dikes and some extensive mineralization.

THE MOON'S IGNEOUS ROCKS

The mountains of the moon look like craters, and there are no folded mountains, features which may be significant in the study of the structures of the earth. There is no evidence of erosion, no atmosphere, no

great ocean. The craters slope nearly 50 degrees like rock slides, probably a result of spalling of rock from temperature changes. The day is a month long and very hot at noon, very cold at night. Some have suggested that the craters resulted from planetesimal bombardment, but spectroscopic study favors rather a true volcanism with rocks that are apparently pumiceous. Astronomers are watching for changes in craters—it is uncertain if any of them have notably changed in 50 years.



FIG. 138.—Some surface features of the moon. (Photograph by Yerkes Observatory.) In the center are large craters with prominent rays, possibly sulphur emanations along cracks. The forms of the common craters show best at the left. In the dark plains (maria) there seem to be relics of old craters almost melted into the plains, in contrast with some sharp younger craters built on the plains and in the rims of the old craters.

There are also great plains in which *craters seem to be melted down!*¹ Small craters of later origin break the plains, but the old plains seem to be smoothed by heat of *magma below*, possibly a zone of molten material near the surface (Fig. 138). No continental drifting on a *sima* or *magma* seems to have occurred on the moon, as indicated by the lack of folded ridges. Yet the *magma* layer must have been extensive near the surface to permit the slumping of the old craters, and tidal stresses must have operated as on the earth to cause the drifting. The absence of folded mountains indicates that earth folds formed some other way.

PHYSICAL CHEMISTRY OF MAGMAS

STATES OF MATTER IN MAGMAS AND RELATED TO MAGMAS

Matter exists in three main states—solid, liquid, and gaseous, but at high temperatures many substances show no sharp distinction between liquid and gas.

¹ BARRELL, Geologic bearing of the moon's surficial features, *Am. Jour. Sci.*, 5th ser., vol. 13, pp. 306-314, 1927.

The temperature and pressure of the maximum on the curve in broken lines in Fig. 139 are called the "critical" temperature and pressure.

The critical temperature of water is 374°C., but the critical temperature is raised by material in solution. The greater the quantity dissolved, the greater is the displacement. At moderate temperatures, silicates are not very soluble, but at 400° they may be extremely soluble—we do not know. At more than 100°C. above critical temperature of water certain liquid glasses may hold 20 per cent water. There seems to be a continuous series between hot solutions of water in glass and water solutions of silicates. In solutions of the less soluble silicates part of the series probably reaches supercritical conditions. Experimental difficulties and the possibility of colloidal conditions in silicates leave the exact conditions in a magma uncertain.

Another change of state, less generally recognized but none the less well known to mineralogists and metallurgists, is a transformation of solids with temperature. Recall, for example, the several states of sulphur. The "geologic thermometer" is based in large part on the recognition of such transformations. Pressure makes a slight difference in the temperature of the transition, but the data give at least a rough guide to the past temperatures of minerals.

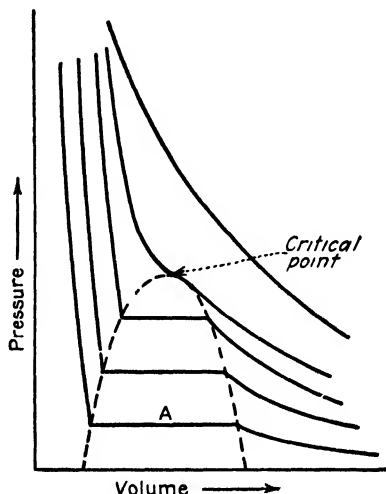


FIG. 139.—Diagram of the critical temperature and pressure. Pressure-volume curves for several temperatures. At A the horizontal parts of the curves show a large volume change while the liquid changes to gas. No such horizontal part appears in the curves for high temperatures.

MAGMAS ARE FUNDAMENTALLY LIQUIDS

Magmas are essentially liquids, though they react with solids and gases and cool to solids. The physicochemical study of magmas is based primarily on the properties of liquids. It has long been recognized that a magma is a *solution* and obeys the laws controlling solutions in general.

In laboratory work on water solutions there is seen a notable tendency for the *solutions to become homogeneous by diffusion* through the solution at a rate depending partly on viscosity. The pressures on a magma where it accumulates are so different, however, from those of simple experiments that some modification of the homogeneity is to be expected even where the liquid phase is the only one present. The base of a batholithic

magma may be under thousands of atmospheres more pressure than its top. Moreover, the rate of diffusion has been estimated as very slow in magmas; possibly 8 cm. of rock may deposit by diffusion from a 7-meter layer of magma in 250 years.¹

Common notions of solutions also distinguish a *solvent* from the *thing dissolved*; for example, salt may be dissolved in water. The distinction becomes less definite when water and alcohol are found to dissolve in each other in all proportions, and at high temperatures it is found that even salt and water may form fluid mixtures in all proportions. The distinction therefore breaks down. Solutions are mutual solutions, and magmas are complex mutual solutions.

Finally, it is characteristic of water solutions that substances *may be dissociated* in them. The molecules of salts in solution are so broken down that the parts may migrate conducting electricity. Melts and magmas dissociate dissolved minerals in much the same way. The melts of enstatite, for example, must contain something besides enstatite molecules, because the first crystals to form when an enstatite melt cools are not enstatite but olivine. Similarly, when melted orthoclase is cooled, the first crystals to form are leucite.

Thus the analogy between magmas and other solutions is very extensive and fundamental. The deep magma, however, is hardly to be compared with the material of a controlled laboratory experiment, because the walls react upon it and receive emanations from it, probably during its whole evolution as a magma.

MAGMAS DISSOLVE GASES

The gas law is briefly stated as $PV = RT$; that is, the pressure times the volume is equal to a constant times the temperature. The law of solution of gas in a liquid is similar.² The solubility varies with the pressure; usually also an increase of heat tends to drive the gas out of solution (but see page 213).

At the surface lava temperatures are about 1000°C. (= 1273° abs.), and pressure is a little more than 1,000 grams per square centimeter. At a depth of 100 km. the pressure is about 30,000,000 grams per square centimeter, and the temperature is perhaps near 1200°C. (= 1473° abs.). A great deal more gas might be held in magmas than the almost negligible quantity that remains in molten rock at atmospheric pressure. The pressure has increased thirty thousand times while the counteracting temperature rose less than 20 per cent. This, of course, does not indicate that greater volumes of gases are present, but only that if present they

¹ BOWEN, N. L., Diffusion in silicate melts, Jour. Geology, vol. 29, p. 315, 1921.

² FROLICH and others, Solubilities of gases in liquids at high pressures, Ind. Engineering Chem., vol. 23, pp. 548-550, 1931.

would be easily held in solution. The rise of pressure dominates the situation, enormously overbalancing the rise in temperature.¹

The gases dissolved in magmas are known as *mineralizers* from their effect in aiding crystallization and developing minerals not otherwise formed. Water gas is the chief mineralizer.

"Pneumatolysis" is a term used to describe reactions due to gases from a magma acting on the adjacent rocks or on the border phases of the igneous mass itself. It is best restricted at present to surface effects near volcanoes. If the term pneumatolysis is used for contact effects it implies that, in certain cases at least, the emanations from a magma at depth are known to be gaseous, not liquid. This is probable but not well verified experimentally (page 219).

There are many reasons to believe that gases are present in magmas, and many features are logically attributed to the action of these gases. Minerals containing water and other mineralizers are known to have formed in deep-seated magmatic rocks. Still other minerals, such as quartz and hornblende, although not containing much mineralizer material, have not been formed artificially by dry fusion without the aid of fluxes. The characters of igneous quartz grains show that some crystallized at temperatures as low as 575°C., and there are few fluxes in magmas except water that could account for such a reduction in the temperature of crystallization. Mirolitic and amygdaloidal cavities result from the separation of mineralizers. Minor explosions in connection with eruptions of lava seem to be closely related to gas, and the larger violent explosions are so uniformly accompanied by gases and rain that they are believed to be related to magmatic gas. There are visible gas emanations from lavas in crater lakes, and near the crater are fumaroles, the visible smoky emanations; solfataras, sulphurous emanations at slightly lower temperatures; and mofettes of CO₂ at low temperatures. Water vapor emanates from Mount Etna near the top even after nine months of dry weather. Geysers and hot springs seem to be later phases of similar emanations. Pegmatites and hydrothermal alteration and ore deposits are also logically related to the concentration of water in magmas. The coarse grain of many pegmatites indicates a growth from a liquid of low viscosity, and the gaseous constituents are those most likely to reduce the viscosity of a liquid (see page 196). Many contact effects include the formation of mineralizer minerals and additions of constituents that are carried in water solutions. Pitchstones contain much water. Even those igneous rocks that reveal no hydrous or mineralizer minerals contain gases, probably in microscopic cavities. Finally, a certain linear type of variation in a series of igneous rocks may logically be attributed to a progressive escape of volatile constituents from the magma. Naturally at the time of escape of gases from solution several of their internal

¹ SHEPHERD and MERWIN, in Jour. Geology, vol. 35, p. 114, compute the gas content to saturate a lava at 1 km., on different assumptions, as between 75 and 1,800 cc. per gram of lava.

effects on the magma are reversed—the crystallization is hastened, the viscosity is increased and some early hydrous minerals may change deuterically into anhydrous minerals. Aplites may result from the escape of gases from a pegmatite magma. See the list of further readings on mineralizers and emanations.

The content of water in normal magma is not certain, possibly not uniform, and we have no device for measuring it. Average igneous rock contains 1.15 per cent water (page 128), and it may be estimated roughly that a little more escapes than is retained when a magma solidifies. Average lava may have 1 to 4 per cent of water, quantities equivalent to more than 70 *volumes of gas* per 1 volume of rock. Detailed work on the magma at Kilauea indicates 4 per cent.¹ Pitchstones seem to have much more than normal magmas, up to 10 per cent or more. The content is probably small in basaltic or primary magma and greater in granitic magma, as is indicated by the more abundant miarolitic cavities in granites and pegmatites, by the ore deposits near the more siliceous rocks, by the series of ferromagnesian minerals—pyroxene and olivine in the gabbros, followed by hornblende in the diorites and biotite and muscovite in granites—and by the laboratory experiments showing that water keeps a silicate melt liquid to lower temperatures, and the fact that granites are known to be later than associated diabases.

Computation of the amount of gas escaping from magmas at Katmai showed that the steam contained about 0.117 per cent HCl, 0.032 per cent HF, and 0.029 per cent H₂S. These figures mean that in 1919 the Valley of Ten Thousand Smokes gave off 1,250,000 tons of HCl!² It is easy to account for the chlorides of the sea by periods of active volcanism.

The probable sources of gas in magmas are two—primary in planetesimals and secondary in additions. It has been suggested that sea water soaks into magmas through the walls, which may act as semipermeable membranes, but the idea has little to support it. It has been shown that capillarity does not serve to introduce water into magmas, and that the often-cited Daubree experiment has no bearing. There is, however, a mechanism involving vapor pressures (Fig. 140) by which water may sometimes enter a magma. Besides these *possible* methods it is well understood that magmas may acquire gases by assimilating or dissolving, either in place or in fragments, some wall or roof rock that contained gas or water. This is “resurgent water.” It is not generally believed that the heat of an intrusive magma will cause the water in an adjacent rock to be volatilized with enough pressure to drive much of that gas into solution in the magma. In the first place, most rocks are porous

¹ TYRRELL, G. W., “Principles of Petrology,” p. 47. See also Econ. Geology, vol. 9, p. 398, 1914; Jour. Geology, vol. 30, pp. 659–672, 1922; and Am. Jour. Sci., 5th ser., vol. 22, pp. 481–502, 1931.

² ZIES, E. G., The Valley of Ten Thousand Smokes, Nat. Geog. Soc. Katmai Ser., vol. 1, No. 4, pp. 4 and 78, 1929.

enough to allow the gas to escape in the other direction to places of lower pressure. If the pores are filled, as they rarely would be, with water under hydrostatic head, this pressure has to overcome the vapor pressure of water in the magma before it enters. The vapor pressure of a magma with several per cent of water in it should be high enough to overcome the hydrostatic pressure of several hundred feet of a water column. Higher pressures of water or of non-porous rock may drive water into solution in the magma.

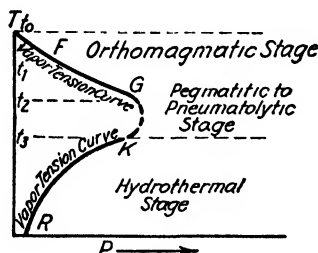
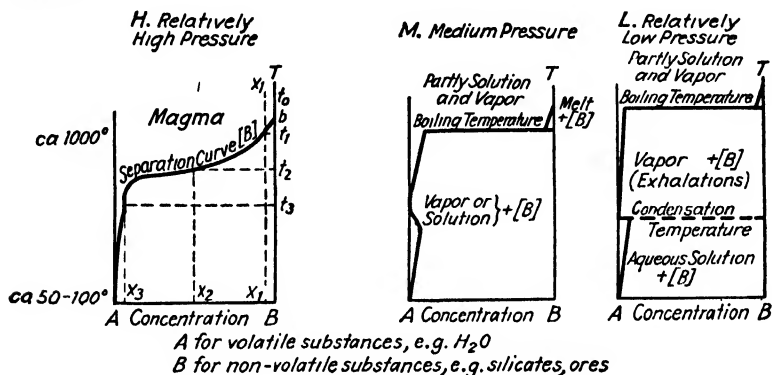


FIG. 140.—Curves of the behavior of melts containing one volatile (A) and one much less volatile (B) component. (Modified from Niggli, "Ore Deposits of Magmatic Origin.") H, Separation curve of magmas with various proportions of (A) and (B) under high pressure; a temperature-concentration diagram. Below curve H is a diagram of temperature and vapor pressure for the same constituents as H. M, The influence of temperature at medium pressure. L, The influence of temperature at low pressure. The external pressure is supposed to be constant in each of the three diagrams H, M and L.

Separation of Gases from Magmas.—Gases may separate from magma solutions for several reasons: (1) change of temperature, (2) release of pressure, or (3) changes in composition, especially the change resulting from crystallization.

The physicochemical relations of magma and gas at different temperatures, pressures and concentrations have been estimated by analogy with other mixtures of volatile and much less volatile components (see Fig. 140).

If pressures are very high, as they must be in the deep reservoirs of magma and deep-seated batholiths, the cooling of the magma may be

indicated by the curve (Fig. 140H), which is the record of a simple melt of two components, one much more volatile than the other. With a small per cent of the volatile component *A*, such as is characteristic of primary magmas, the temperature of crystallization of the non-volatile *B* is reduced from t_0 to t_1 . The growth of crystals of the non-volatile component changes the composition of the melt along a curve, falling rapidly at first. The flatter part of the curve means that at about t_2 crystallization proceeds rapidly and becomes almost complete. This is the orthotectic stage (page 45). The residual solution is rich in the volatile component, but there is no break in series to the most dilute solutions.

Corresponding to this curve of crystallization, the vapor pressure of such a melt shows remarkable changes (see the lower curve of Fig. 140H). Through the orthotectic stage the pressure rises in spite of the reduced temperature, because the residual liquid has an increased proportion of the volatile component. When the concentration of volatile material reaches a high figure, perhaps 40 to 80 per cent, the pressure reaches a maximum, after which cooling reduces the vapor pressures. The maximum, in magmas consisting of silicates and water in large part, *may carry the residual melt above critical conditions*. Recall that the critical temperature of water is 374° , and that at such temperatures water dissolves very little of rock-forming minerals. There is a sharp disagreement among physical chemists as to whether the hot hydrous residue of magma may decompose the silicates of a normal rock magma, so as to produce soluble alkali salts in quantities great enough to make a noteworthy modification in the critical temperature (compare Figs. 140 and 147). In any case, the hydrous magma has its viscosity much reduced, the temperatures are probably about 500° to 600°C . and the vapor pressures are very great, but there is still considerable concentration of non-volatile matter. This is the pegmatitic stage. Niggli and others use also the term "pneumatolytic" assuming that the magma reaches a gaseous state (page 213). Further cooling brings the hydrothermal stage with more watery solutions, largely in the liquid state.

The high vapor pressure of the magma in the pegmatitic stage makes this magma especially likely to force its way out into cracks as emanations, into the schistose wall rocks. Although this magma may be gaseous in the sense of being above critical conditions, yet it resembles a liquid in both density and concentration of silicates.

Special note should be made of the maximum in the curve of vapor pressure in Fig. 140. This means that more gas can be dissolved in the magma at very high temperatures than at pegmatite temperatures. If a very hot magma is saturated with water vapor, a cooling would so increase the vapor pressure as to tend very strongly to drive gases out of solution. Very high pressures would be needed to hold them in solution at the pegmatite stage. This driving of gas out of solution by cooling is

strikingly the opposite of the familiar laboratory experiment in which gases are driven out of solution by heat. This ordinary trend is illustrated by the residual magma after it passes the pegmatitic stage, when any increase of temperature tends to increase the vapor pressure and drive gas out of solution.

Consider next the effect of reduced pressure during the orthotectic stage. If the pressure falls below the curve *FG* of Fig. 140, the magma boils, producing a gas rich in the volatile component (see Figs. 140*M* and 140*L*). At the same time the non-volatile component crystallizes in quantity. The changes here are complex and cannot be predicted too definitely in systems so complex as magmas. Some of the gases that boil out are probably very dilute. Others may be formed above their critical temperatures and be rich in the non-volatile components. Such solutions should very closely resemble the pegmatitic magma shown by the curves of Fig. 140*H*, but they differ very notably in the conditions of formation—one by cooling, the other by reduction of pressure.

The cooling history of magmas at reduced pressures is shown by Figs. 140*M* and 140*L* in contrast to the curve of Fig. 140*H* for high pressures. At low pressures, such as those of lava flows, the gases largely boil out at the high temperature of the magma at the time of extrusion. Some bubbles may be trapped in the lava as vesicles, and if the cooling is so rapid as to form glass some of the gas in solution may not have time to separate. At medium pressures, such as characterize laccoliths, sills, stocks and plateau batholiths, part of the gas boils out and forms vesicular sills or causes contact action around the intrusive. The pressure may hold enough gas in solution, however, so that the magma follows later a course much like that in Fig. 140*H*.

It is noteworthy also that the curves have certain retrograde portions, indicating some resorption during the pegmatitic and hydrothermal stages.

A reduction of the confining pressure during the pegmatitic or the hydrothermal stage is probably less significant. At the pegmatitic stage the magma residue is already much like a gas (possibly above critical conditions), and if so it could not boil to produce a gas. At the hydrothermal stage the liquid can boil, but it is a very dilute solution and the gas that boils out will not be very different in composition from the solution.

The outline given above is complicated by the peculiar nature of pressures on magmas at depth. If water fills the pores of relatively large openings in the roof, its pressure may be hydrostatic; but if the roof is wholly impervious, the whole weight of the roof exerts a hydrostatic pressure. Neither case is to be expected. The common roof rocks have a very small porosity, and the magmatic solutions exert their vapor pressures on the few small pore openings as well as on the rock load. Therefore, in spite of some frictional resistance in the capillary and subcapillary openings, there will be a slow leakage of gaseous emanations. Even a

deep-seated magma under a heavy covering of rock may give off gases, if the cover is fractured or porous. The small openings may allow gases to penetrate into regions of less pressure even when the viscous magma as a whole does not enter the openings. At a depth of 3 km. a magma with a viscosity of 1 would penetrate rock openings 0.001 mm. in diameter very, very slowly. Gases from magmas probably penetrate rapidly even these small openings and may impregnate the wall rocks with silicates.

The final example of gases escaping from a magma is that which results from assimilation. The magma may assimilate some rock that disturbs the solubility of the gas. This is quite aside from the evident fact that the assimilated rock may have contained such an abundance of gases as to furnish some of the emanation. The action may also be that of a catalyst, but this case is probably not important.

It is noteworthy that the two chief modes of separation of gas from a deep-seated magma, one by reduction of pressure and the other by change of composition during crystallization, should occur at different stages of magma history, one early and the other late. The early emanations have been called "exhalations" to distinguish them from late "residues." There is much difference of opinion as to the relative importance of the two sorts of emanation. Probably the slow emanations by boiling out into porous wall rocks are practically continuous from an early to a late stage of magma history.

It is a question how far pressures on a magma really are reduced allowing the escape of emanations. The confining pressure on a liquid with an impervious cover is not the rock column above but an inverted *cone* of rock above. The pressures may thus be surprisingly high on emanations from magmas near the surface.

Nature of Magmatic Emanations. A. The direct method of attack is that on gases from lavas. Where all precautions are observed, no rare gases like the argon of the atmosphere are found in the emanations, so that the gases are probably free from atmospheric contamination. Water is very largely predominant, in many gases 80 per cent, and in some 98 per cent.

The work of Brun in Europe casts doubt on the presence of water in magmas, both as to its detection in the emanations and as to its possible existence at the temperatures of basaltic lavas. Day and Shepherd, however, *found it* and are sure of their results (see the list of further readings). It is recognized, of course, that ground water near a volcano may cause explosions.

Following water are CO_2 , N, and sulphur gases, commonly in the order given but varying with location and conditions. In the condensed water are also F, Cl, NH_3 , and some of the common bases of igneous rocks, Na, Fe, etc. Commonly H_2S is more abundant than SO_2 . A long series of other elements and compounds are reported here and there, and copper, arsenic, and other metals such as form ores have been detected in the deposits. Fumaroles are commonly acid in reaction.

The gases are not in equilibrium as they escape but react with each other. Some heat may be generated by the reacting gases, and more is generated as the gases burn in the presence of air. Expansion due to relief of pressure may cause some cooling, but this is probably subordinate.

Emanations vary in a single crater in a single day in both amount and kind. *The more gas the hotter the lava*, though it is uncertain which is the cause of the other.

Naturally the emanations near a volcano may vary with the distance they have traveled through rock, the kind of rock and the temperature of the rock. Many fumaroles also have no doubt been contaminated with air.

The fumarolic deposits are largely made up of halogen compounds of NH_4 and the common metals of rocks.

B. A second method of study of emanations is experimental with silicate melts. Although the composition of the liquid and vapor phases are the same at the temperature of boiling under certain pressures, both liquid and vapor are probably highly aqueous when the vapor boils out (see Fig. 140). The solution would probably be dilute.

C. Finally, a great deal of information about emanations is obtained from pegmatites (pages 69 to 73) and contact deposits (pages 421 to 424); and most students of ore deposits believe that even hydrothermal action should be attributed largely to emanations (page 425). Tourmaline and other "mineralizer" minerals introduced into rocks are quite generally accepted as signs that some part of that introduced material emanated from a magma.

The conclusive proof of large additions by contact action refers back to classic studies by Lindgren at Clifton-Morenci, Arizona, and by Kemp at Tamaulipas, Mexico. It required careful analyses of the original limestone and the altered rock, in addition to good evidence that the volume had not been greatly changed, to convince petrologists generally that in contact ore deposits the main effect was that of addition rather than a leaching and recrystallization.

It also became necessary to show that the solutions were truly magmatic emanations rather than simply heated ground waters. Umpleby found this very clear at White Knob, Idaho, and it is now agreed that many contact rocks contain essential contributions from magma.

The common contact emanations are silica, iron oxide, soda, and other prominent elements of the magma, apparently carried in large part by water and mineralizers. The mineral andradite is produced from limestone, albite forms in shale, and many other products are noted. Although silica, iron and sodium are not abundant in the gaseous emanations around volcanoes, it is noteworthy that traces of them have been recorded in many fumarolic deposits.

Magmas of different composition give off different emanations. The surface emanations studied are largely from basalts. Those from other surface rocks are supposed to be similar. The deeper, basic magmas yield little contact emanation as compared to more silicic magmas. Some

chlorine may be added. The relation of ores to different magmas, described on page 141, is also significant.

If it is granted that large volumes of material may emanate from magmas, a question remains as to the process of separation, and whether the emanation is gaseous or liquid. If contact emanations are liquid, the whole discussion is out of place under the present heading of gases in magmas, but it is left here because of an impression that even a liquid emanation may be largely governed by its gas content. The physics of gas separation is outlined above (pages 213 to 216). Can as good reason be found for a liquid to separate from a magma? The answer is, Probably it can.

If a magma contains a variety of constituents, some of which are of much lower molecular weight than others, those of low weight may migrate by some osmotic or other effect to the places of low pressure in the magma, usually therefore toward the top (see Fig. 141). Such a migration involves chiefly the gaseous, light constitu-

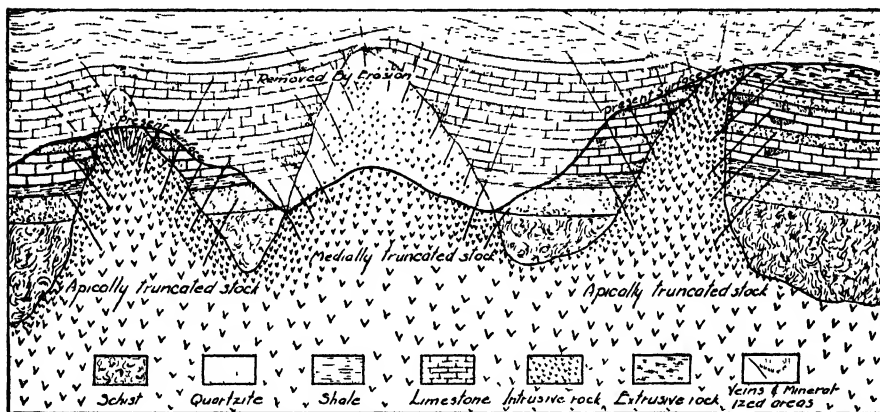


FIG. 141.—Ideal section through the top portion of an intrusive body. (After Butler, *Econ. Geology*, vol. 10, p. 119, 1915.) The lighter portions of the stocks are differentiated into the more siliceous bodies, and the solutions from the magma deposited metals mostly around the apex of the upward projections of igneous rock.

ents and is referred to as a "streaming of gases,"¹ as if a quite different and more rapid effect than diffusion. This is not clearly explained in most discussions of streaming gases. It is possible that a slight migration by diffusion causes a difference in density in the magma, which then moves more rapidly in convection; but this is based on an assumption that gases dissolved in a magma reduce its density. Experimental data are much needed. NH_3 dissolved in water makes it lighter, but HCl in the water makes it heavier. We do not know how gases affect the density of magma.

Certainly if gases streamed over toward one part of the magma chamber, they would not only make the magma different in different parts, but would probably greatly reduce the viscosity of the magma where the gases became abundant, permitting much more penetration of the openings in the wall rock than would be possible for the main viscous magma. This "advance guard" of the magma would be notably different in composition from the main mass and would constitute a real emanation from the chamber. It therefore seems possible that there may be either liquid or gaseous emanations or both. There is still active discussion over the ques-

¹ DALY, R. A., "Igneous Rocks and Their Origin," p. 276, 1914. PAUL NIGGLI, "Die leichtflüchtigen Bestandtheile in Magma," Leipzig, 1920. G. W. TYRRELL, "Principles of Petrology," p. 160.

tion whether or not such liquid emanations are probable in contact action. Recall the uncertainty as to whether critical conditions are reached (page 214), and the suggestion that magmas themselves may replace some wall rocks (page 188).

The gaseous nature of contact emanations is perhaps most strongly indicated by their pervasive penetration of the rocks, in contrast to the relatively sharp-walled fissures of deep veins. The propulsive force necessary to drive a liquid magmatic separate into its wall is obscure, but the change of state when gas separates from solution involves a large expansion. This is not wholly conclusive, as it is possible that if two liquids separate one would have such a low viscosity that it would penetrate submicroscopic openings.

The physical chemists also feel certain that emanations must be *gaseous* because of their acidity, observed at volcanoes, and explained by the hydrolysis of sulphide and halide emanations. These acid emanations could hardly arise from a *liquid* emanating from a magma in which the alkali silicates predominate largely over halides and sulphides; only the emanations have the concentration of sulphur and halogens.

Opposed to the idea of the gaseous nature of contact emanations is the occurrence of associated tin and tungsten ores, supposedly brought in as gaseous fluorides. It is known, however, that one fluoride boils at 19° and the other at 705°C., so that a gaseous condition would tend to separate them widely.

Opposed to the idea of deep-seated acid emanations are the observed alterations of rocks near contact ore deposits. The most common acids that enter into the composition of the minerals of contact deposits are the weak acids H_2S , CO_2 , and SiO_2 , and there are strong alkalies like soda and potash, which would make the combination alkaline in its effects on rocks. The common sericitization and introduction of sulphides indicate alkaline solutions, for sericite does not seem to be formed commonly by acid surficial mine waters, and sulphides are carried by alkaline solutions. Clearly there is some difference between the deep emanations and fumarole gases, though perhaps not so great a difference as that between the points of view of economic geologists and vulcanologists. Lacroix has found evidence near Vesuvius of some emanations near the surface much like deep-seated pegmatites, so that he emphasizes the continuity of the process.¹ Some heavy metals such as form ores at depth are detected at fumaroles. The halogen compounds of many metals can be carried as gas or in liquid solutions either alkaline or acid. It should be added that the emanations may be so hot as to be dissociated and perhaps neither acid nor alkaline. The formation of sulphides may be a late reaction during cooling rather than a precipitation from a hot alkaline liquid.

It is probable that both gases and liquids emanate from magmas. Certainly the field evidence is strong that two *fluids* separate during deep magmatic action. The question whether at such depths they are two liquids or a liquid and a gas, as at the surface, may well be left to the study of physical chemists. If the chief emanation is a gas, it is dense

¹ LACROIX, A., Étude minéralogique des produits silicates de l'éruption du Vésuve (April, 1906), Nouvelles Archives du Muséum d'Histoire Naturelle, 4th ser., 1st Fascicule, 1907.

enough to dissolve silicates almost like a liquid; if it is liquid it is hot enough to penetrate dense rocks about as easily as a gas. Possibly early emanations are acid and gaseous, and the late emanations are more alkaline and liquid.¹ Any late mother liquor, filter-pressed or squeezed out from its interstitial condition to constitute a liquid emanation, is quite certain from its origin to be alkaline.

Sequence of Emanations.—Aside from the theoretical sequence of emanations due to different processes—release of pressure, cooling (or heating), and crystallization—which may produce somewhat different gases, observation shows a variation of emanations with temperature and the stage of igneous action. At Vesuvius the observations show at about 1000° alkali chlorides, HCl, CO₂, and SO₂; from 1000° down to 400°, HCl and SO₂; from 400° to 350°, NH₄Cl, (NH₄)₂CO₃, NH₃, and HCl; below 350°, H₂S, and finally, CO₂. The sequence is not exactly the same at other volcanoes, but nearly all show changes. At Katmai the chlorine, fluorine, and sulphur, and their hydrogen compounds, acids, are found in gases around 600°, whereas the fumaroles at 100° have only traces of them.

In emanations from deep-seated sources the sequence can be inferred from the mineralogic alterations, which indicate that early and hot alterations, above 1000°C., form silicates; around 1000°C. and below, come iron oxides; later the sulphides are deposited; and at about 750°C. some emanations deposit cassiterite. Some contact ores no doubt formed at about 575°C. or less.² Possibly there may be a series of effects in the first group, garnet, epidote, and vesuvianite indicating greater heat and earlier formation than wollastonite and diopside.

Gases and Pegmatites.—The descriptions of pegmatites emphasize the signs of an abundance of water and other gases in the pegmatite magma. The methods by which gases become so concentrated should now be clear. Two main ideas have been advocated as to the origin of pegmatites and both may be true.

The *prevailing notion* is that the pegmatite magma is a mother liquor left after much of the magma has crystallized as anhydrous minerals. The vapor pressure rises to a maximum during cooling and crystallization and the water content rises to 50 and even to 90 per cent (Fig. 140H, pages 213 to 216). The residue may be "filter pressed" from the partly crystalline mush into crevices formed by the shrinkage from cooling of the outer shell of igneous rock or into fissures resulting from dilatancy (page 195). The *other notion* is that the thinly fluid emanations from a really liquid magma may escape as pressure is reduced (Fig. 140M, page 215) or may penetrate in advance of the magma itself into crevices where the

¹ ZIES, E. G., Valley of Ten Thousand Smokes, Nat. Geog. Soc., Katmai ser. vol. 1, No. 4, favors acid emanations in depth as well as at the surface.

² VOGT, J. H. L., Magmas and igneous ore deposits, Econ. Geology, vol. 21, p. 330, 1926. Vogt estimates higher temperatures than most petrographers.

viscous, less hydrous magma would go only slowly, and not far before solidifying (pages 215 and 216).

Both of these ideas have much to support them. The similarity of the rocks formed by early emanation and by late "filter pressing" is attributable to the fact that both emanations would be rich in water. There may be no really sharp distinction. By analogy with contact metamorphism, it seems that emanations may begin early, but continue until late in magmatic action, and the volatile emanating materials, no doubt act as agents of transportation of minerals, possibly including most pegmatitic minerals. The results of differentiation also suggest that the expulsion of mineralizers and their dissolved load of minerals continues from an early hot stage to a late cooler more hydrous stage (page 211) See list of further readings.

The evidences of early emanation of some pegmatites are that they are found largely in magma walls and borders of granites; some are crossed by later pegmatites; many are altered by later passing fluids; siliceous introductions precede ores in contact action; silicic stringers have been found below a gabbro, in which early basal segregations were rich in olivine; and it is known that gases can penetrate where viscous magma does not.

The coarseness of crystallization is commonly explained by the growth of crystals from a passing solution; this would also explain the possibility of a concentration of some rare element in a large single crystal. The thin fluid nature of the magma, like a hot-water solution, would also explain rapid diffusion of material to a growing crystal.

It has been suggested that aplite and pegmatite magmas form and accumulate in essentially similar ways, but that in some places water may escape from such a magma, and the relatively dry magma forms aplite; if the water does not escape the wet magma forms pegmatite.

Are pegmatites igneous or aqueous rocks? There was for many years a question whether pegmatites were veins or dikes. Spurr has recently discussed them as "vein-dikes" or veins formed from "thick magma" more or less hydrous and gelatinous. Pegmatite magma has many analogies with the fluids forming most ores near igneous rocks. The analogy is evident to almost everyone, but there is some discussion as to the content of water in pegmatite and ore magmas, also as to whether they merge gradually into hot-water solutions or are to be contrasted with such solutions as concentrated liquids contrast with dilute. The truly igneous character of the pegmatite magma is perhaps best indicated by the field fact that late magmatic or deuteric effects are in pegmatites quite similar to those in granite itself. Physical chemists agree that there may be a continuous series of stages from orthotectic to pegmatitic conditions (pages 213 to 215). It is noteworthy, however, that the series is also continuous beyond the pegmatites to the hydrothermal stage. There is no gap between magmas and water solutions under the conditions in which pegmatites form. The continuity is broken where pressure is reduced, say where a magma is intruded within a few hundred feet of the surface,¹ and it is noteworthy that peg-

¹Spurr estimates 500 to 1,000 feet, but this depth must vary widely with the structure of the rock (see Geol. Soc. America Bull., vol. 36, pp. 568-569, 1925).

matites do not seem to have formed at shallow depths; the effects near the surface are hydrothermal even up to the hot springs. It is customary also to distinguish even at depth those deposits formed from dilute hydrothermal solutions as veins and those from more-concentrated magma as dikes, in spite of the continuity of conditions. For most petrographers the evidence of attack by hot water at depth is just as convincing as that of such action near the surface, and the deposits are classed as aqueous rather than magmatic, even though the magmatic origin of the water is clearly recognized. Probably a large majority of petrographers include the intermediate pegmatitic stage with the orthotectic stage under the term "magmatic" and thus include pegmatites as igneous rocks.

MAGMAS DISSOLVE LIQUIDS

It is a question whether magmas dissolve other liquids, water being a gas at magmatic temperatures. It was early suggested, however, that

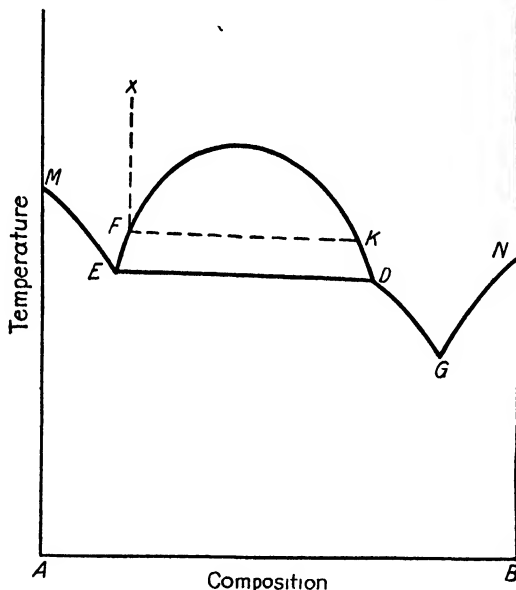


FIG. 142.—Diagram of partial miscibility. When a mixture with a composition indicated by XF cools to the temperature F , the melt separates into two liquids. On further cooling one liquid follows the composition indicated by FE ; the other the composition KD . The two liquids may differ in gravity enough to cause a separation of two layers, but the separation may be incomplete. The later history resembles that of a eutectic (Fig. 146).

two main magmas, basaltic and granitic, might mix or dissolve each other to form intermediate rocks, and an exposure has been described in which the field evidence suggests mixed or hybrid magmas.¹

Immiscibility and Partial Miscibility.—Liquid solutions of liquids are of several kinds. Alcohol and water mix in all proportions, but oil and

¹ HARKER, ALFRED, Tertiary igneous rocks of Skye, Geol. Survey United Kingdom Mem., pp. 177–194, 1904. The term "hybrid," however has also been used for magma with contaminations by any solid rock

water dissolve very little in each other, and other liquids have solubilities that are limited and that vary with temperature—"partial or limited miscibility." For example, phenol and water mix in all proportions at temperatures above 250°C. but tend to separate into two layers on cooling, one layer containing a little water in phenol and the other a little phenol in water (see Fig. 142 for a generalized diagram).

Melts of silicates with sulphides of iron separate on cooling into two liquid layers, as Vogt long ago showed by a study of slags. It is very probable that such sulphides as those at Sudbury, Ontario, separated as a liquid from solution in magma (Fig. 143). Pyrite, having a melting point

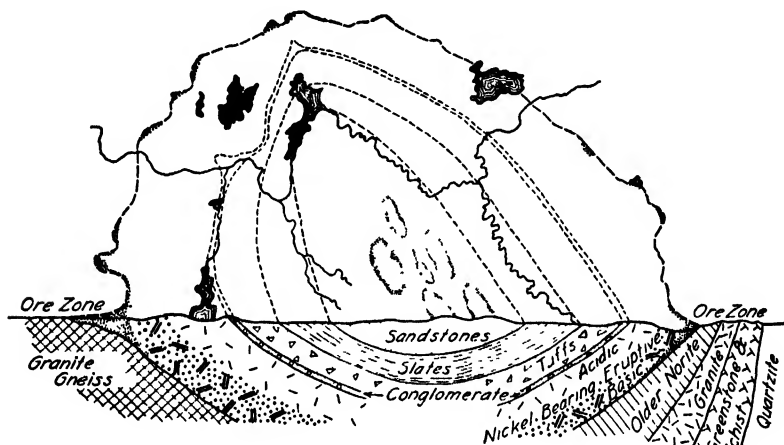


FIG. 143.—The Sudbury, Ontario, intrusive. (After Coleman, *Jour. Geology*, vol. 15, pp. 761-763, 1907.)

at high pressure above that of silicates, may segregate by early crystallization, but pyrrhotite, if segregated early, would separate by immiscibility because no crystals would form at the high temperature of basic magmas. More-ordinary silicate melts, however, have failed to show such separations in laboratory tests. To be sure, the melts of lime and silica, and those of magnesia and silica do show a limited miscibility (Figs. 149 and 153) but this is in a range of composition not closely approached by magmas, a range high in silica.¹

The field and laboratory criteria of unmixing are not satisfactory. Those physical chemists who insist that the only criterion should be the finding of glass globules in glass overlook certain factors. Natural glasses may cool too fast to permit the separation. On the other hand, where cooling is slow the separated globules ultimately crystallize and no doubt lose their globular form, as the sulphides have done at Sudbury.

¹ KRACEK, F. C., The cristobalite liquidus in the alkali oxide-silica systems, *Am. Chem. Soc. Jour.*, vol. 52, pp. 1436-1442, 1930.

CRITERIA SUGGESTING THE UNMIXING OF MAGMA LIQUIDS

Note in the first place that no natural occurrence of unmixing of magmas has ever been conclusively proved.

Proof may be found in the occurrence of sharply separated glassy globules in a natural glass; especially if the globules are uniform in size and distribution; especially if the globules are loosely attached indicating a composition and shrinkage different from the matrix; especially if the matrix has flow structure or foliation but the foliation does not affect the globules.

Suggestions are also found in:

1. Rocks with a ratio of silica to bases greater than 3:1.
2. Sulphides in an igneous rock, especially in masses accumulated near the base.
3. Differentiated masses with sharp contacts between differentiates in approximately gravitative arrangement.
4. Sharp banding in igneous rocks.
5. The simultaneous extrusion or intrusion of two or more magmas of different compositions. For example, late complementary dikes.
6. Orbicular structures without foreign cores.

This leaves a good deal of uncertainty as to whether the separation of partly miscible magma fluids is of any importance except where sulphides are present. The arguments are many but inconclusive. The field evidence is clear that when pegmatites are formed, and probably when other emanations issue from deep-seated magma, *something* separates from the fluid magma, and the emanation is richer in mineralizers than the remaining magma. The emanation *may* be gaseous, but it produces igneous rocks so often as to suggest a liquid separate, as was long ago suggested by Arrhenius. The question of immiscibility is complicated by the fact that a liquid can yield two different liquids by *streaming of dissolved gases* (page 218), without any true immiscibility.

Other suggestive features are as follows: The flows from a single volcanic outlet may be rhyolite and basalt, in an alternation repeated so many times in a single geologic period, that it seems they must have come from two layers, both liquid at the same time. Regional variation in magma supplies has been taken to mean a simultaneous existence of different liquids under the region. The occurrence of two kinds of differentiated series in a single chamber is suggestive (Fig. 157).

Some men still find the idea improbable for rock magmas, but it should be retained as a working hypothesis. The separation may be related to the *water* content; not many silicate melts of high water content have been tried. (See the list of further readings.)

MAGMAS DISSOLVE SOLIDS

The commonest notion of solution is that of a solution of a solid in a liquid, and the most familiar solution is that of salt in water. Magmas are solutions at high temperatures, and the distinction of solvent and solute is, to say the least, obscure, if it can be made at all. The tendency is to consider that thing as solvent which remains liquid longest when the

solution is cooled; in most magmas this is the siliceous, hydrous, possibly pegmatitic, residual magma. Quartz, feldspars and mineralizers in magma become saturated with apatite, zircon and ores early in the cooling. The analogy with water solutions is very rough, however, and many factors modify the process of crystallization.

The solid most commonly available for solution in a magma is the wall rock or a fragment from the wall or roof.

Assimilation.—There is no doubt that magmas may assimilate or dissolve their walls to some extent and under some conditions. The generation of magma in the depths of the earth is perhaps at first a process of fusion, but it becomes largely a matter of assimilation as the volume of liquid grows. The term assimilation is not usually defined so as to include such magma generation but is used to describe any solution of walls, roof, or fragments *in the lithosphere* or the *reservoir of magma*. There is much question whether assimilation is a significant feature in the upper lithosphere, where the results may become accessible to our observation. Still the zone of magma generation is not so sharply separated from the lithosphere that we should expect any abrupt change in the capacity of a magma to dissolve its walls. No doubt assimilation is fairly extensive high into the lithosphere. It is clearly best not to describe the effects along contacts as "fusion." Rocks modified by such assimilation are called *contaminated* or *hybrid* igneous rocks. *Endomorphic contact action* is in large part a matter of assimilation, but endomorphism includes textural effects and changes due to the emanation of volatile constituents.

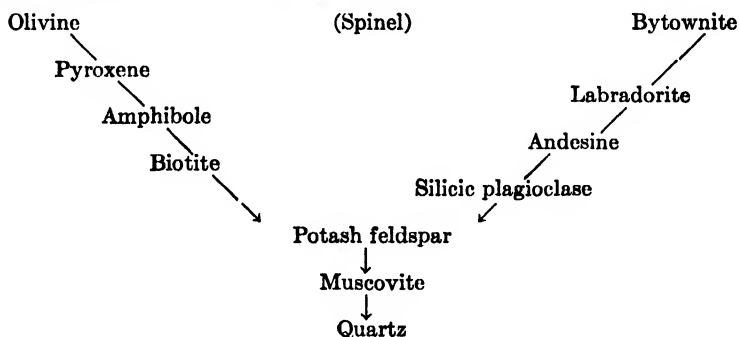
Two kinds of assimilation may be distinguished, marginal and abyssal. The first implies a local solution of the wall of the magma chamber, but abyssal assimilation implies a stoping of the roof and walls into fragments, which may settle to great depths before solution. The theory of abyssal assimilation removes the problem from the zone of directly observable effects; the probability may only be inferred from general considerations and from more remote effects. Only the marginal assimilation leaves its record in the rocks as an observable rounding or solution of fragments or a relation in composition between the intrusive and the country rock.

In the zone of observation we find plenty of evidences that magmas dissolve rocks. See the criteria of assimilation. Craters of inactive volcanoes are often cooled to solid rock, and as some volcanoes become rejuvenated, the rocks are assimilated rather than being blown out. Blocks of wall rock are rounded by solution, and the magmatic rocks near the walls are enriched in constituents of the wall (Figs. 27 and 28). The temperature of many lavas is high enough to melt some common sediments, like shales, many of which flow at 1000°C. (see also the note, page 132, about the effects of inclusions).

An argument based on the above observations may be considered as additional evidence. The inclusions found near the roof and walls of a batholith were probably formed just before the magma solidified. As they are somewhat attacked and assimilated by the magma at this late stage, the earlier hotter magma must have assimilated even more actively.

Minerals of Assimilation.—When an igneous magma assimilates an igneous rock, the minerals that result should not be commonly abnormal in any way, but when it assimilates sediment or altered sediment, the hybrid nature of the resulting material may be very striking. The minerals staurolite, sillimanite, and cordierite in igneous rocks are strong indications of contamination. Very commonly near limestones the magmas contaminated with lime develop diopside and other lime silicates, and a few such hybrids seem to contain nephelite. The minerals and rocks that melt below 1300°C. are those which might be expected to be readily absorbed in the magma.

Bowen¹ has made a study of the mineral changes to be expected in inclusions in a magma. Anticipating the study of separation of solids from solution we have the suggestion that magmas give *series* of reaction minerals as follows (page 242)



A true assimilation or solution should occur only when a magma at a certain level in the chart includes fragments of minerals listed lower in the series, or some less common mineral of similar chemical reaction. If a mineral higher in the series was included, the magma might cause it to recrystallize or might react with it to form such a mineral as was crystallizing at the time. This distinction between reaction and solution is very necessary, for the results may not always be easily distinguished. A basalt (labradorite-pyroxene) should not dissolve in a crystallizing granite magma but might react to form a biotitic gneiss with more siliceous feldspar.² On the other hand, granite fragments in a crystallizing basalt magma might dissolve and leave no trace whatever.

¹ BOWEN, N. L., Jour. Geology, vol. 30, p. 513, August suppl. 1922.

² FENNER, C. N., The crystallization of basalts, Am. Jour. Sci., 5th ser., vol. 18, p. 251, notes that rhyolites in Yellowstone Park, as if unaware of this restriction, melted basalt inclusions (see U. S. Geol. Survey Mon. 32, Part 2, pp. 430-432, 433, 1899.)

It is to be noted that the common sedimentary minerals calcite and kaolinite are not members of the series, and their behavior in inclusions is not so definitely determined.

These reactions are not incompatible with a suggestion based on field and analytical data that there may often be partial assimilation by an *exchange* of material,¹ the magma and the xenolith each contributing some constituents to the other. Possibly alumina and potash tend to migrate from the xenolith to the magma at the same time that lime and magnesia are added to the xenolith. The mechanics of the exchange is obscure, but this need not raise any doubt as to the fact of the exchange. The difference in fusibilities of minerals in a xenolith is sufficient to make a selective attack almost certain.

Conditions Favoring Assimilation.—The conditions that favor assimilation are: large volumes of magma, especially deep masses, with some superheat; hydrous or other mineralizer content to delay solidification; some turbulence or movements in the magma chamber to bring fresh magma supplies; and finally a kind of wall rock that is soluble in the magma concerned. Several of these factors, especially the superheat, have been subjects of uncertainty to petrographers.

Question of Heat and Solution.—To get a wall rock into solution in a magma usually requires heat, first for raising the temperature to the melting point and then for supplying the latent heat of fusion. What heat is available?

1. The observed superheat in basaltic magmas may be 200°C. or more but is probably not often more than 100°C. This superheat of about 100° covers all the sources of heat involved in the generation of magma (page 145). This is enough to heat up and fuse only a small per cent of wall rock. The superheat to be expected in a magma depends a good deal on what theory is adopted as to the evolution of a magma. If it is assumed that nearly all magmas evolve from basaltic magma by crystallization (pages 244 and 245), it is hard to see that much assimilation can occur in any but the primary magma. The other magmas would not exist except as they lost their superheat and crystallized in part. While they crystallize it is unlikely that they dissolve very much.

It has been suggested that the heat needed to dissolve a crystal is about equal to the heat that it liberates when it again crystallizes from solution. Thus it is said that the heat of solution is not all used up in solution, it is used only temporarily in small amounts to get the crystal into solution long enough to let it crystallize out again as an igneous mineral, possibly of different composition;² the crystallization restores most of the heat used in solution, allowing further solution. This is not clear. The crystal grows mostly as a result of cooling and this can not restore any heat

The field evidence of assimilation in granites is so strong that it may not only show the presence of some superheat in granite magma but may even serve as evidence that granite magma may form in some other way

¹ READ, H. H., *Petrology of the Arnage district*, Geol. Soc. London Quart. Jour., vol. 79, pp. 479-484, 1923.

² BAIN, G. W., *Evidence of assimilation and assimilation processes*, Jour. Geology, vol. 34, p. 661, 1926.

than by crystallization and separation from basalt magma. The hydrous magmas probably assimilate their walls even down to 570°C.¹ If gases "streamed" (page 218) to the upper part of their chambers, the modified gas-rich magma might actively flux its roof.

2. This leads to a second factor, namely, that the solution of a rock with mineralizers in it may so lower the temperature of fusion of the resulting mixed magma as to be equivalent to adding heat. Contact heating may also add water to magmas, with similar effects, but this action is believed to be rare (page 212).

3. Dissolving a crystal in a magma may produce heat of solution and heat of reaction. For example, the solution of CaCl_2 in water gives heat. On the contrary, the solution of NH_4Cl in water absorbs heat. The reaction of an acid and a base to form a salt usually evolves heat, so that "acid" rocks acting on "basic" magma might be expected to give heat. "Basic" magma, however has about 50 per cent silica, so it is not really very likely that heat is evolved as magmas dissolve rocks.

4. The heat for solution may come from a rise of isotherms—a transfer of deeper heat—rather than an evolution of new heat.

Evidences That Assimilation Is Not Commonly Important.—The arguments against the importance of assimilation are not based wholly on the heat problem just outlined. Rocks are known in which differentiation has clearly produced a series grading from a large bulk of gabbro to a smaller, later mass of granite, and the inclusions in the granite are of such a nature that if they had been largely assimilated they would have made the magma more like the gabbro from which it was derived. Again, some of the largest known intrusive masses have along their floors enormous volumes of xenoliths large and small, much metamorphosed but not assimilated. Although there was no doubt some assimilation, it was preceded by a metamorphism that made the rocks much like the enclosing magma, and assimilation stopped while the masses of undissolved included material were still very great. It seems likely then that abyssal assimilation depends largely on the size, and the rate of settling and rate of solution of xenoliths. If large rock fragments sink into a batholith they may sink so deep before dissolving that no portion of them will ever again see the light of day. On the contrary, if many small fragments settle a little and are dissolved, they may furnish a good deal of material to the upper differentiates of the batholith, which are the parts accessible to us for study. The rates of settling are wholly unknown. Many blocks falling into batholiths from their roofs, however, are known to be enormous, Fig. 144. If the larger blocks constitute as large a percentage as seems likely, their assimilation probably occurs only at depths comparable to that of the asthenosphere. Finally it may be shown that, in

¹ ESKOLA P., On contact phenomena between gneiss and limestone in Massachusetts, Jour. Geology, vol. 30, p. 292, 1922.

many places where the structural relations might indicate favorable conditions for assimilation, there has been none to speak of. The igneous mass shows no approach in composition toward that of the wall.

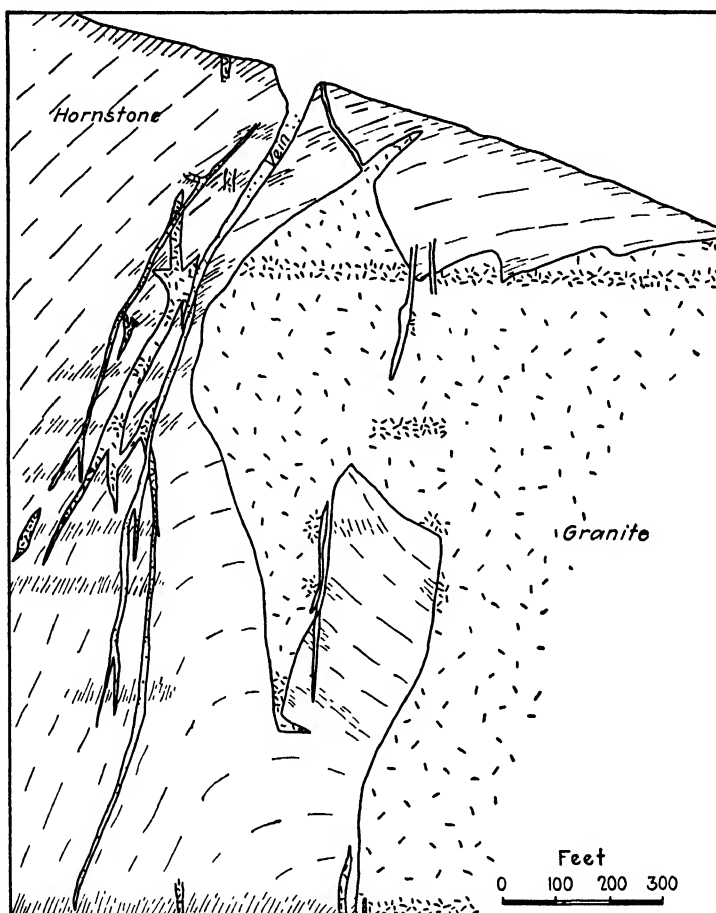


FIG. 144.—Vertical section, Drumlummon Mine, Marysville, Montana. (After Barrell.) Large blocks of hornstone seem to have been caught in the process of being stoped off into the granite.

Conclusions as to Assimilation.—There are many rocks which show clear evidence of assimilation, and many others in which the evidence is equally clear that practically no assimilation has occurred. The doubtful cases are so numerous, however, as to be the basis of much speculation. The question of the importance of assimilation, in a broad way, hinges largely upon the rate of sinking of stoped blocks and the rate of their solution in the magma. The abundance of unassimilated shale and basalt (hornfels) xenoliths of great size in the basal parts of some of the largest known intrusive masses makes it probable that abyssal assimila-

tion is so deep and so slow that the products of this action are not likely ever again to reach the light of day. Some small fragments are no doubt dissolved and may affect the course of differentiation. The supply of heat needed for solution of the solid walls and fragments is a problem of importance in smaller intrusions and seems to limit the assimilation to be expected to a fraction of the volume of the original magma. (See the list of further readings.)

CRITERIA SUGGESTING ASSIMILATION

1. Granitoid rocks, with no chilled borders, are most favorable to assimilation.
2. Rounding at edges of included fragments, and schlieren.
3. Projections of certain wall rocks into intrusive, alternating with embayments in others, indicate assimilation of embayed rocks.
4. *Lit-par-lit* injection, and schlieren near inclusions.
5. Dark-colored patches in igneous rocks suggest assimilation of xenoliths; some weather differently from the main rock.
6. A narrow zone near the border of the intrusive of different composition from the main mass.
7. Gradations in composition from the walls or inclusions to the including rock or intrusive.
8. Thin film of glass in an adjacent rock or fragment along the contact of an igneous rock.
9. Orbicular structure—concentric shells of different composition around centers, which may or may not exhibit a xenolithic nucleus.
10. Presence of assimilation minerals in fresh igneous rock; staurolite, andalusite, cordierite, calcite, wollastonite, sillimanite, zoisite, graphite, scapolite or melilite. Less definitely garnet, corundum, nephelite, epidote, large volumes of biotite, amphiboles or pyroxenes.
11. Some minerals in abnormal association, as quartz in a gabbro or garnet in a granite.
12. Abnormal chemical composition (as, for example, over 20 per cent CaO), especially if enriched with an element abundant in the adjacent wall.
13. Blending of two igneous rocks.
14. Much contact action by an intrusive suggests it is capable of assimilating.

Syntexis is a term used to imply that a magma has been formed by the melting of more than one kind of rock, or that a primary magma has assimilated some solid rock. The assimilated material may or may not affect the later differentiation. Some petrologists have suggested that many of the observed cases of differentiation have been caused by the assimilation of rocks containing water or carbon dioxide.

Syntexis has been appealed to in explaining how the more siliceous and the more alkalic rock clans can be derived from primary basaltic magma. It is clear that weathering and sedimentation divide uniform igneous rocks into at least three things as different as sands, shales and limestones; if now one of these kinds of rock is assimilated, profound changes in the magma will result.

A general criticism of the idea of the syntexis is possible, in that it makes no attempt to explain the whole series of igneous rock clans, but only those which our common sediments might produce by assimilation. The series of rocks that have not been so explained—peridotites, pyroxenites, ores etc.—are just as truly part of a continuous series as basalt and granite, so that it is not clear why a wholly different explanation of their origin should be favored. On the other hand, the alternative view of differentiation by crystallization, reaction, settling and filterpressing explains the whole series. It may be admitted, however, that some rare sediments might by syntexis yield these rocks; for example, in some gabbros the ore bodies, not commonly explained by syntexis, might be of this origin.

A particularly strong, persistent effort to prove the importance of syntexis has been made on the basis of alkalic rocks. Jensen, in Australia, suggested an assimilation of salt; others the syntexis of limestone or the reaction of pegmatitic emanations and limestone. As a concentration of mineralizers may produce alkalic rocks and a simple cooling reaction will give leucite + quartz \rightleftharpoons orthoclase, it is evident that syntexis is not a vital necessity.

The discussion of syntexis hinges largely upon the probability of important assimilation, and the evidences against important assimilation in the lithosphere (page 228) are also against much syntexis.

Formation of Solids from Magma Liquids. *Crystallization.*—Solids result from magmatic liquids as a result of cooling and possibly at times as a result of change of pressure and of escape of gases. The solids may later react with the liquid to form other solids and change the liquid. The process of solidification is usually complex, and the discussion must progress from simple to complex features.

A simple melt of one component on cooling some 40° to 60° below the temperature of crystallization or "saturation" will begin to crystallize. The equilibrium between the melt and the crystal suspended in it is supposed to be kinetic, not static. If a crystal remains of constant size, it is because molecules are added as fast as they are dissolved in the melt. Crystals grow from a melt with the elimination of the latent heat of fusion; that is, if the melt is left to radiate its heat, the temperature curve is regular until crystallization starts, then remains at constant temperature until the mineral is crystalline, and then again falls (Fig. 145). The cooling may be rapid in small intrusives and flows, but calculations based on the known poor conductivity of rocks show that large bodies of magma must have taken hundreds of thousands of years to cool. Leaving out of account all convection and the heat effects of crystallization and emanation, it is calculated that it takes 6,400 years to cool a point 700 meters inside a magma chamber with its walls originally at 0°, from a primary temperature of 1000°C. down to 400°C.¹

Glasses.—If a melt is cooled rapidly, crystals may fail to get started until the temperature falls perhaps 120° below the temperature of crystallization, and the melt may become very viscous—supercooled or supersaturated. Viscosity varies in magmas even above the temperature of

¹ GRUBENMANN, ULRICH, and PAUL NIGGLI, "Die Gesteinsmetamorphose," p. 250.

crystallization according to composition. High silica makes magma highly viscous, and water content reduces viscosity. In nature we get glasses, with a viscosity that is practically infinite and a rigidity higher than that of many crystals.

In the behavior of matter under deforming stresses rigidity should be carefully distinguished from strength. Rigidity is shown by a return movement on release of stress; strength by the amount of stress required to effect continuous deformation. Some writers define solids in terms of rigidity and some in terms of strength. If the substance lacks both it is fluid, if it has both it is solid, but some substances have only one. Glasses are solids, whereas magmas are liquid. The curve of viscosity with temperature is apparently continuous, but it changes direction very notably and increases from 10 to $(10)^{18}$ poises in a certain temperature interval; for common glasses

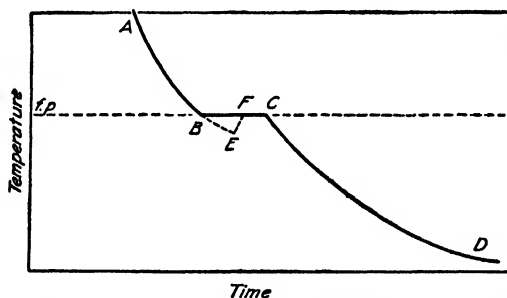


FIG. 145.—Cooling curve of a melt that crystallizes at the temperature BC.

this interval is between 550° and 1200°C . After this steep rise it is probably no longer advisable to consider glass a liquid, not even a supercooled liquid.¹ Several other physical properties change abruptly in the region of abrupt change in viscosity.

If a slowly cooling magma has in it some small crystals of the minerals, as "seeds" to start growth, there is very little supersaturation. Most of the large deep-seated magmas may be seeded by xenocrysts from the walls or inclusions. In the absence of seeding some magma intrusives may be greatly supercooled, and if they are not too viscous when crystallization starts crystals will grow rapidly in all directions from a center of growth, forming spherulites.² This extra rapid growth continues until the liberated latent heat raises the temperature of the adjacent melt up to the normal crystallization point, Fig. 145. Spherulites, as well as perlitic cracks and lithophysae which form only in glasses, are proof that some rocks that are now crystalline were once glasses. Apparently even the infinite viscosity of glasses will not prevent new

¹ HAWKES, L., On rock glass, *Geol. Mag.*, vol. 67, pp. 17-24, 1930. EITEL, *Phys. Chemie der Silikate*, especially Fig. 94. But see also WILLIS, "Geologic Structures," 2d ed., p. 434, 1929. Prolonged stress "within the elastic limit" is said to deform glass but not crystalline solids.

² If molten menthol cools on a slide, the spherulites can be watched in process of growth under the microscope.

crystallization if metamorphism takes place; and so glasses become devitrified. We find no Archean glasses and few that are pre-Carboniferous.

There is also probably a relation between the increasing viscosity of a magma and the formation of a porphyritic texture. Large crystals grow in a fluid magma, and the groundmass of fine crystals probably forms only after the cooling and possibly the loss of mineralizers has made the interstitial residual magma more viscous. It is also suggested that high viscosity in a slowly cooling magma may cause the formation of orbicular structures.¹ The viscosity and rate of diffusion of crystallizing laboratory melts were found by Bowen² and indicate at least roughly that magmas have viscosities from 4 to 200, several hundred times as great as water.

Rate of Cooling and Size of Grain.—Size of grain that results from cooling under conditions of conduction is proportional to

$$\frac{T_1 - T_2}{U_1 - U_2}$$

where $T_1 - T_2$ is the time of cooling and $U_1 - U_2$ is the drop in temperature. For example, suppose a magma which would crystallize at 1000° is intruded at 1200° into a rock at 200°. The margin at once comes to a temperature of 700°, which is below the point of crystallization. Its grain size (ignoring any circulation of magma) should be zero; it is glassy. Glassy borders form on many small intrusions.

The temperature a few inches inside the borders of the intrusive would fall much more slowly as heat is conducted through the glass into the wall rock, probably reaching 1000° so slowly that small crystals would form. Farther in from the borders still slower cooling would produce coarser crystals, and in the centers of large intrusives uniform slow cooling gives uniform granitoid textures.

Some intrusives show coarse grain out to the borders. Two cases are possible: (1) If the magma used as an example in the preceding paragraph was intruded into rock that had a temperature of 900°, the first adjustment would bring the border down to 1050°, not cool enough to crystallize. Then the slow conduction of heat into the wall would give coarse crystals even at the border. In other words, if the temperature of the magma is farther above its temperature of crystallization than that of the wall is below, there will be no reduction of grain size at the border. (2) If a large magma in contact with a cold wall is locally cooled enough to start convection, the circulating magma will be bringing fresh supplies of heat to the wall until part of the wall itself may become highly heated and even assimilated, and the heat conduction is very slow.

¹ SEDERHOLM, J. J., On orbicular granites, Finland Geol. Comm. Bull., No. 83, pp. 60-77, 1928.

² BOWEN, N. L., Crystallization-differentiation in silicate liquids, Am Jour. Sci., 4th ser., vol. 39, p. 186. 1915.

In Michigan lava flows, there are several applications of the study.

1. Amygdaloid inclusions can be distinguished from upper zones of flows even in drill cores.
2. The maximum grain shows the thickness of a flow.
3. Abnormally slow change of grain indicates a drill hole oblique to the structure.
4. The change of grain shows which is the way to the nearest amygdaloid.
5. Sudden change at a joint shows displacement.
6. One may calculate from changes how far it is to the next amygdaloid.
7. One may distinguish boulders from bed rock.
8. Flows are never coarse out to the border; intrusives may be. (See the list of further readings on grain size.)

Equilibrium Diagrams. *Eutectics.*—The crystallization of mixtures of two components which form no compound with each other may be

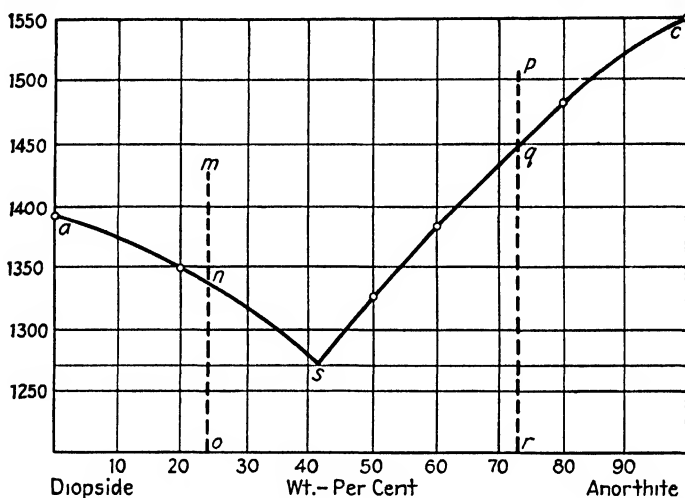


FIG. 146.—Equilibrium diagram of the system diopside-anorthite. (Modified from Bowen.) See text for the significance of the figure. A simple eutectic of two rock-forming minerals.

illustrated by the curve for diopside-anorthite mixtures (Fig. 146). Any proportion of these constituents may be represented by a point on the base line. The temperatures are shown by vertical distances from the base line, and the melting point of diopside is *a*, that of anorthite is *c*. Consider the composition indicated by the line *mno*. At a temperature *m* the two components are in mutual solution and liquid. As cooling proceeds, a temperature *n* will be reached at which the diopside will begin to crystallize. Similarly, starting from *p*, a temperature *q* will be reached at which anorthite begins to crystallize.

By a series of experiments the loci of all points such as *n* and *q* can be determined. This is the curve *asc*. Along *ans* diopside will crystallize; along *cqs* anorthite will crystallize; at *s* both crystallize together as a *eutectic mixture*.¹ The temperature of *s* is the *eutectic point or temperature*.

¹ The term eutectic is from the Greek and means "melt easily."

The cooling history may be illustrated by starting at *m*. First the solution cools until at *n* it becomes saturated with diopside. Then diopside crystallizes, and as a result the composition of the liquid changes as it cools to *s*. At that temperature both diopside and anorthite crystallize until all is solid. After that the crystalline mass cools to lower temperatures.

An exactly similar history results from a mixture *p*, except that anorthite crystallizes first.

There may be several such eutectic mixtures that are significant in the development of igneous rocks, but nearly all rocks are more complex, and the simple eutectics probably never develop in natural magmas. Several suggested "rest magmas" are very questionable.

The intergrowth of quartz 27.5 per cent and orthoclase 72.5 per cent is probably more often attributed to the eutectic process than any other, but there are serious objections to this view. Neither mineral melts as such. Orthoclase melts incongruently (Fig. 150), and quartz may be transformed before melting. If the graphic mixture was a true eutectic, it should be expected in most granites, but it is not found in most granites, only a granitoid intergrowth is found in which orthoclase is formed earlier than quartz. The graphic structures seen in quartz-feldspar phenocrysts in a glassy groundmass are certainly not late products of eutectic origin but appear to have formed early in the cooling of the magma.

TABLE XI.—SEVERAL EUTECTICS OF ROCK MINERALS

Minerals and Their Melting Points		Eutectic Temperature
Diopside, 88 per cent, and 1391°	Forsterite, 12 per cent 1890°	1386° ± 3°
Diopside, 84 per cent, and 1391°	Tridymite, 16 per cent 1670° ± 10°	1362°
Albite, 72 per cent, and 1100°	Tridymite, 28 per cent 1670° ± 10°	About 975°
Diopside, 3 per cent, and 1391°	Albite, 97 per cent 1100°	About 1085°
Diopside, 61.5 per cent, and 1391°	Leucite, 38.5 per cent 1686° ± 5°	1300° ± 2°

Morey determined a eutectic between water and potassium nitrate (Fig. 147) on a curve which differs notably from the curve (Fig. 140) for a volatile and a non-volatile component. The latter is probably applicable to magmas, because the magmatic silicates are much less soluble than alkali silicates.

Eutectic systems with compounds of the two components are a little more complex, and some such compounds occur in magmas. The compound is located by a maximum in the curve of temperatures of fusion (Fig. 148). Between the compound and each component there is a eutectic. The complexity is increased if the components form compounds with several different compositions. Lime and silica form as many as five compounds. These melts, like the simple eutectic, may be complicated by minerals that are unstable at their melting temperatures.

The binary system MgO-SiO_2 is very important in igneous theory,¹ Fig. 149. There is a simple eutectic between periclase, MgO , and

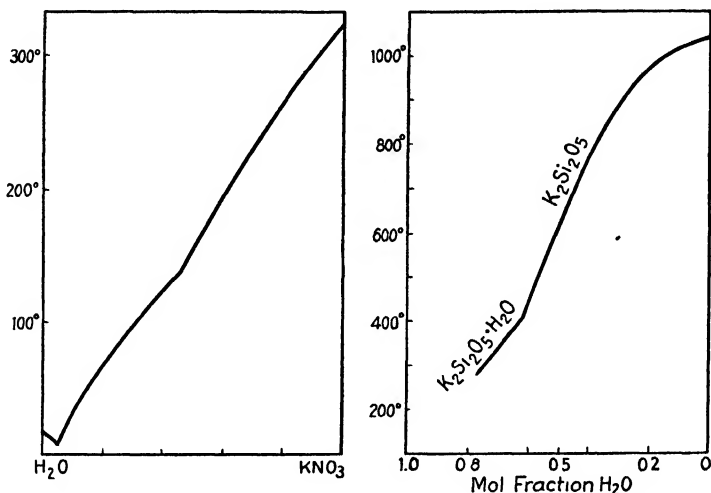


FIG. 147.—Fusion curves of two binary systems involving water. (After Morey, *Engineers Club of Philadelphia*.)

forsterite, Mg_2SiO_4 ; but the rest of the series is complicated by a transformation of clinoenstatite MgSiO_3 , and by limited miscibility near the silica end of the series. It is especially noteworthy that *there is no eutectic* between the olivine and the pyroxene. The curves show that a melt with the composition of enstatite on cooling will yield first crystals of olivine. On further cooling *these react with the melt to form pyroxene*. This principle of reaction and change of minerals is a more important feature in magmas than the simple eutectic (see page 242).

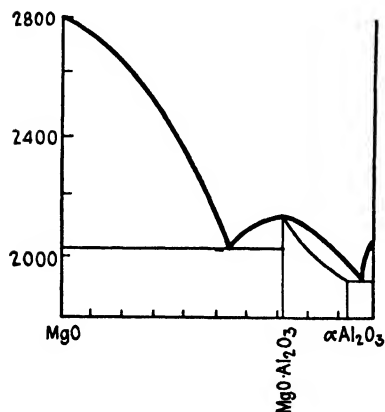


FIG. 148.—Equilibrium diagram of the system $\text{MgO-Al}_2\text{O}_3$. (After Rankin and Merwin, *Jour. Am. Chem. Soc.*, vol. 38, p. 571, 1916.) An example of a binary system with a compound—spinel. The equilibrium between spinel and alpha-corundum has some complexity.

Three different rates of cooling in the system MgO-SiO_2 may give different results. Rapid cooling of a melt of pyroxene might carry the melt below the reaction temperature before crystallization began; pyroxene

crystals would then grow in the supercooled melt. Moderately

¹ ANDERSON and BOWEN, *Am. Jour. Sci.*, 4th ser., vol. 37, pp. 487-500, 1914. GREIG, *Am. Jour. Sci.*, vol. 13, 5th ser., vol. 13, pp. 15-17, 1927.

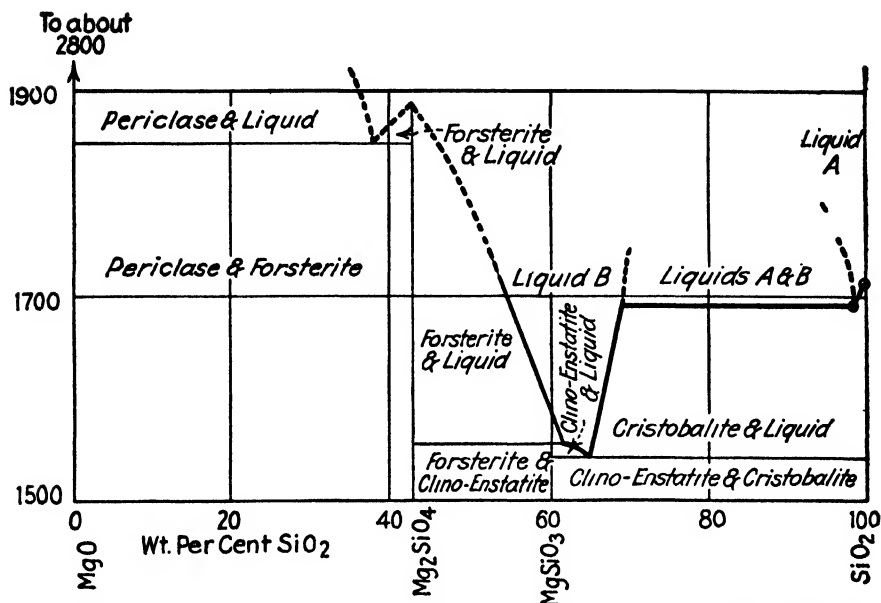


FIG. 149.—Equilibrium diagram of the binary system MgO-SiO_2 . (After Bowen and Anderson; and Greig.) See the text for notes on this system.

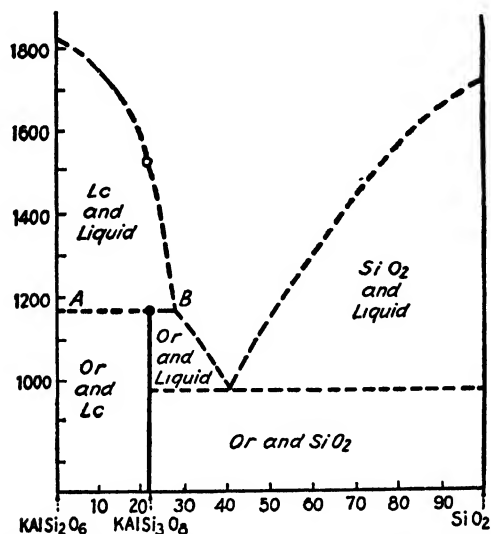


FIG. 150.—Diagram suggesting the sort of equilibrium believed to exist in the system leucite-silica; incongruent melting of orthoclase. (After Morey and Bowen, *Am. Jour. Sci.*, 5th ser., vol. 4, p. 10, 1922.)

rapid cooling might result in some olivine crystals; these, at the reaction temperature, would tend to react with the melt to form pyroxene; but if the time was short, the reaction might be incomplete leaving some early corroded olivine surrounded by later pyroxene and quartz. Very slow cooling would form olivine which, at the reaction temperature, would all turn to pyroxene, unless it was separated from the melt (pages 244 and 245).

A somewhat similar reaction occurs in the common potassium aluminum silicates (Fig. 150). A melt with the composition of orthoclase, when cooled so that crystallization begins, will yield crystals of leucite. These will on further cooling react with the liquid to form orthoclase. Orthoclase, however, is unstable above 1170°C.

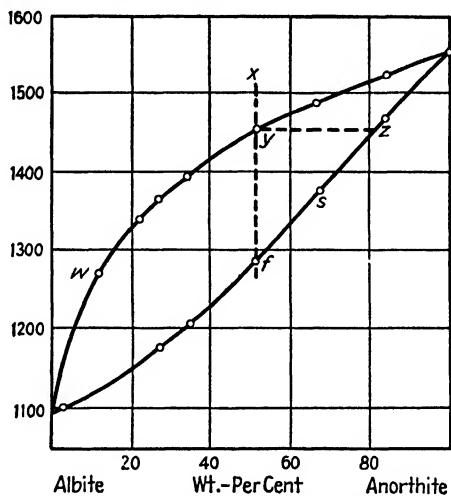


FIG. 151.—Equilibrium diagram for the plagioclase feldspars. (After Bowen.) A common type of curve for solid solutions.

These two laboratory systems show that by simple progressive cooling of a melt, some common mineral may first be caused to crystallize and then to disappear. It is evident that the corrosion of the early minerals of a magma does not necessarily mean a fresh supply of heat.

Solid Solutions.—Isomorphous substances have no ordinary eutectic point. "Solid solutions" and "mix crystals" are terms for the crystals formed from melts of isomorphous mixtures. For these substances the progress of crystallization is usually a "reaction," in this case a *continuous* reaction series, as contrasted with those like the leucite-silica series, which may be called *discontinuous* series. This way of crystallization is illustrated by some of the most abundant igneous minerals.

The curves for plagioclase (Fig. 151) show no minimum except at one extreme. The double curve means that instead of having a melting point the mix crystals have a melting *range*. Three cases arise depending on the rate of cooling, much as in the system MgO-SiO_2 .

1. Rapid cooling of a mixture, shown by the point x , may carry the temperature below the lower curve before crystallization is fairly started. In this case small crystals grow with the composition x .

2. In slower cooling of a melt x , crystals begin to form at a temperature y , but the crystals formed have a composition shown by a point z on the lower curve at the same temperature as the point y . As the crystals grow, the composition of the melt is offset down the slope from y toward w . The crystals growing from a changing liquid show layers of changing composition from z down the slope toward f . In feldspars, for example, the central core or zone is commonly more calcic than the outer zone, and there is no limit to the change till albite forms on the outside.

3. In *very* slow cooling crystallization starts as in slow cooling. The calcic feldspar formed at z , however, when it receives an outer zone, formed at s for example, stands so long in contact with the more sodic melt that the central core, still hot, reacts to attain a composition s , and

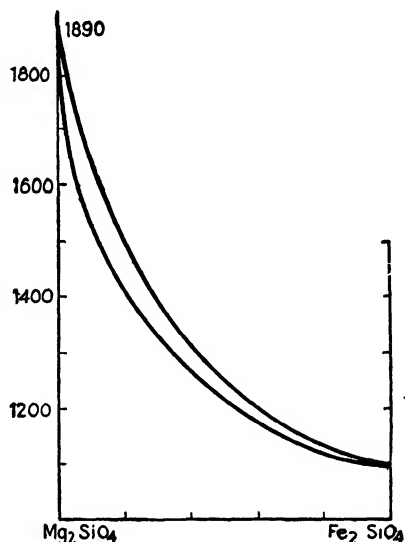


FIG. 152.—The suggested equilibrium diagram for olivine. (After Vogt, *Jour. Geology*, vol. 29, p. 522, 1921.)

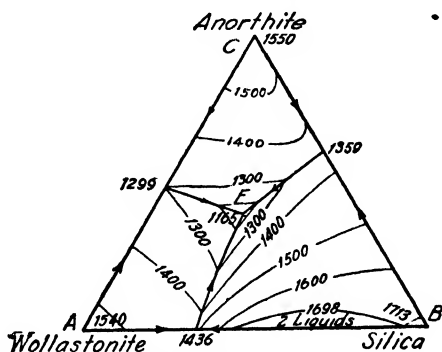
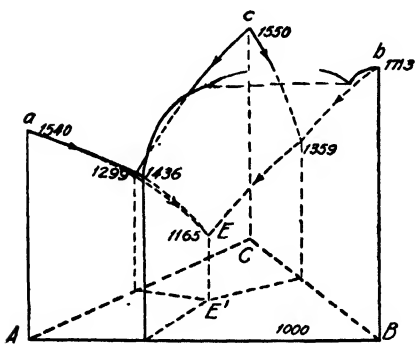


FIG. 153.—Equilibrium in a ternary system of simple character. The arrows on the contour map show the course of crystallization with falling temperatures.

the process of growth and reaction continues until all the crystal has the composition x .

The olivine group probably constitutes a similar isomorphous series (Fig. 152).

More Complex Systems.—Although the reaction series just outlined are fundamental to magmatic theory, each is based on a study of two components, whereas a magma has commonly at least eight or ten. When a three-component system forms a eutectic, the diagram of its relations is a solid figure such as that sketched at the left of Fig. 153. The arrows show the lowering of temperature. To represent such a figure on a plane surface, we ordinarily make a contour map of its upper surface. The system anorthite-wollastonite-silica is an example.¹

In the area near *A* the first crystal to grow will have the composition of *A*. This will change the material of the melt along a line moving directly away from *A*, say toward the boundary curve between *A* and *B*. When the melt has a composition

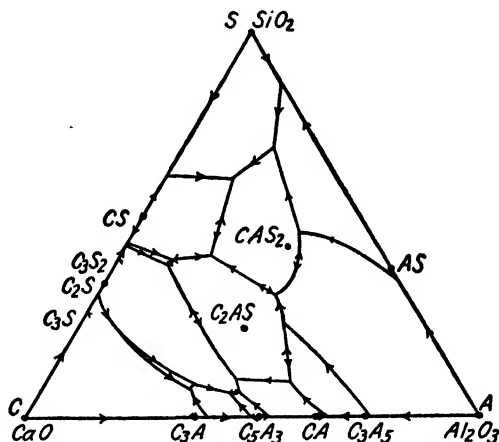


FIG. 154.—Equilibrium diagram to show the complexity of crystallization when ternary solutions of CaO , Al_2O_3 , SiO_2 are cooled. (After Rankin and Wright.)

indicated by that boundary curve, the two constituents *A* and *B* begin to grow simultaneously, and the composition changes toward *E*. After that point is reached, all three components grow together until the mass is solid.

Some ternary systems are complicated by the formation of compounds between two or more of the components. In the simplest case the diagram is doubled. These systems are further complicated if one of the compounds is unstable below its melting point. Such a system is illustrated by anorthite-silica-alumina. Corundum may form early and change the melt until the compound sillimanite or mullite begins to grow, and alumina redissolves.

The system anorthite-silica-forsterite, from which olivine may form early and be corroded as crystals of enstatite grow, is another illustration. It is common to find olivine crystals corroded. Evidently this re-solution of olivine may be due to equilibrium reactions during simple cooling. There is no need of reheating a magma to make it corrode early crystals.

The diagram of the ternary system $\text{CaO-Al}_2\text{O}_3\text{-SiO}_2$ is a fair example of the complexity of a melt of three components, yielding under various conditions some 12 kinds of crystals (Fig. 154).

¹ BOWEN, N. L., "Evolution of Igneous Rocks," p. 39, 1928.

The next degree of complexity is supplied by mix crystals in a ternary system. Such an equilibrium is illustrated by a melt of diopside and plagioclase (Fig. 155). The general course of the melt is either straight away from diopside or slanting up to the left from anorthite toward the central curve, thence moving to the left on the curve toward albite, which is the last mineral to crystallize. Three cases arise here in ternary melts, as in binary melts when mix crystals form. Cooling at a moderate rate gives zoned crystals, but rapid or very slow cooling does not.

The important feature of this discussion of the complexity of magma systems is that we get away from an unduly simple reference to eutectics. Three components provide so many phases that the results are hard to

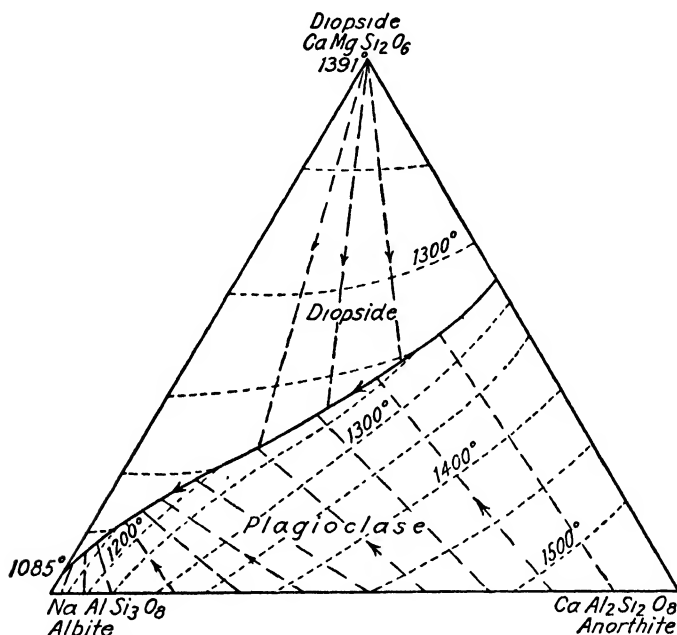


FIG. 155.—Equilibrium in the system diopside-albite-anorthite. (After Bowen.) A ternary system including a solid solution. No eutectic point is reached in this system.

apply in discussions of magma, and the 10 or 12 abundant components of magma may introduce still other complications, some of which may be wholly unsuspected.

Pressure Effects.—In the crystallization diagrams given above, pressure has been ignored. The direct effect of pressure on melting points was noted on pages 152 to 156. Temperature seems to be more important than pressure in crystallization, for it takes several hundred feet of rock to make as much change as 1°C.

It is possible that a change of pressure may affect the order of crystallization and even the character of the minerals that separate from a magma. The materials tested indicate only a slight shift in the eutectic

with pressure (Fig. 156). Certain temperatures of transformation are raised by pressure, that from alpha to beta quartz being perhaps 50°C. higher at high pressure.

Indirectly, also, pressure modifies all these diagrams, perhaps fundamentally, by holding in solution water and gases which may greatly influence the products of cooling.

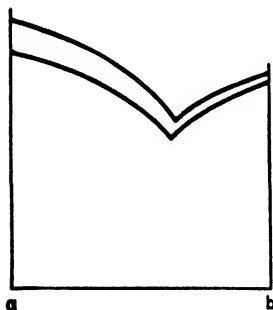
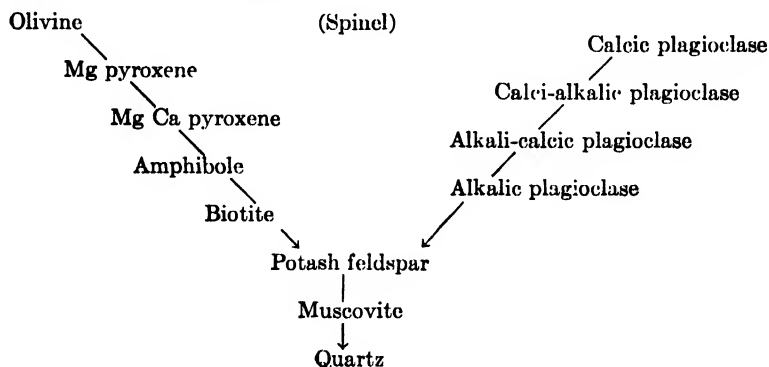


FIG. 156.—The change in a binary eutectic as a result of uniform pressure. (After Vogt, *Jour. Geology*, vol. 30, p. 615, 1922.) Although the temperature of fusion is raised by pressure the composition of the eutectic is only slightly affected.

The effect of pressure on solution or magmatic assimilation is also small, being of the order of 1 per cent per 1,000 atmospheres, but the result is of a different order of magnitude if gas or vapor is involved, as may be judged by analogy with the change in solubility of CaCO_3 if CO_2 is added under pressure (page 333).

Differential pressure during crystallization may develop structures in a crystalline rock resembling that of metamorphic gneisses. Weinschenk refers to piezo-crystallization, but we have no criteria to distinguish it from dynamic action except that it develops early—say before some related aplite dikes.

The Reaction Principle.¹—The equilibrium diagrams already presented exhibit a feature that is believed to be much more important in magmatic action than eutectics. This is the reaction principle. Bowen distinguishes continuous reaction, as in plagioclase, and discontinuous reactions, as in a series from olivine through pyroxene and amphibole to biotite. His generalized series is:



For illustrations of the reaction principle one has only to recall the many rocks in which biotite is altered to magnetite, augite altered to hornblende, and hornblende altered to biotite. Where the altered

¹ BOWEN, N. L., The reaction principle in petrogenesis, *Jour. Geology*, vol. 30, pp. 177-198, 1922; The behavior of inclusions in igneous magmas, *Jour. Geology*, vol. 30, pp. 513-570, 1922.

products are coarse and fresh they probably resulted from magmatic reactions. Recall also zoned feldspars and the corrosion of olivine, quartz and many other minerals. Very commonly a graphic or vermicular intergrowth along the borders of a mineral may be due to reaction of the crystal with a late residue of magma. In gabbros a number of different minerals show such "coronas" or reaction rims; olivine, biotite, enstatite, plagioclase and magnetite may be intergrown in the rims. In more silicic rocks quartz penetrates the edges of feldspars in a similar late fringe, the myrmekite. Even graphic pegmatite may be a late reaction rather than a primary intergrowth.

These late reactions in igneous rocks are described by Sederholm's term "deuteric effects," and the minerals formed by them are called "synantetic."¹ "Paulopost" is a term of similar meaning. The terms are needed to distinguish those alterations that are in direct continuation of the consolidation of the magma of the rock, from secondary changes due to a distinctly later period of alteration. The deuteric effects have special interest because the late magmatic residues may be related to ore deposits and pegmatites. They are most extensive and complex in granitoid rocks.

Most reaction at a late magmatic stage involves a concentration of mineralizers and a development of hydrous minerals. The temperatures are still high, and recrystallization is easy. The new minerals exhibit a remarkable penetrating power; tourmaline penetrates quartz, actinolite penetrates feldspar, quartz and silicic feldspars corrode plagioclase.

NORMAL SEQUENCE AT GARABAL HILL

(After Harker)

	1. Iron ore	2. Olivine	3. Augite (diplage)	4. Brown hornblende	5. Green hornblende	6. Biotite	7. Plagioclase	8. Orthoclase	9. Quartz
a. Olivine diplage rock.	+	+	+	+	+	+	+	+	+
b. Biotite-diorite	+	+	+	+	+	+	+	+	+
c. Hornblende-biotite granite	+	+	+	+	+	+	+	+	+
d. Porphyritic biotite granite.	+	+	+	+	+	+	+	+	+
e. Eurite vein.	+	+	+	+	+	+	+	+	+

The condition of a magma is kinetic rather than static at any time, and equilibrium is easily disturbed by changes of temperature, pressure or concentration, especially by a change in mineralizers, which may be added

¹SEDERHOLM, J. J., On synantetic minerals and related phenomena, Geol. Comm. Finland, Bull. 48, 1916.

by solution or lost by emanation. It is suggested, for example, that a rock that might normally yield an analcite groundmass would probably form albite and nephelite if the water escaped. Biotite also requires water, so that the discontinuous series from olivine to biotite would be greatly modified if there was no increase in the concentration of water. Many magmatic reactions are reversible. In a broad way it may be true also that pressure influences the equilibrium, either directly or by holding mineralizers in solution, so that the evolution of the magma can be logically correlated with its tectonic setting (pages 130 and 131).¹ The early history of a magma must have a very important influence on its later development.

Some late reactions, it must be added, may be reactions in solids still at high temperatures, possibly illustrated by the coronas in common gabbro. There is also the unmixing of a solid solution, probably illustrated by the formation of perthite, as albite separates from solid solution in a potash feldspar. (See the list of further readings.)

Interrupted Reaction Series. Settling and Floating of Crystals.—In general, the crystals that form in a cooling melt are heavier than the melt. In some diabase sills large plagioclase crystals have risen to the top showing that they were lighter than the magma, but this is no way abrogates the general rule. The specific gravity of the whole rock is greater than that of the melt, so an average crystal formed in magma will sink. The formula for the rate of settling spheres given by Stokes² has limited application, but the range of its applicability is discussed by Allen.³

$$x = \frac{2gR^2(d - d_1)}{9V}$$

where x is the rate in centimeters per second, R is the radius in centimeters, d is the density of solid, d_1 is the density of the liquid around it, g is the acceleration of gravity, and V is the viscosity.

In large bodies of magma this tendency of crystals to separate from the melts in which they form entirely breaks up the normal course of crystallization shown by these curves of reaction series.

Take first the plagioclase series. With *very* slow cooling a melt of labradorite might yield a crop of bytownite crystals. If these settled to the bottom, the magma above would then produce labradorite to settle as a next layer. The settling therefore would give differentiation instead of reaction or would modify the reaction.

The effect of sinking crystals is equally important in the discontinuous reaction series. If a melt of enstatite cooled to the beginning of crystalli-

¹ NIGGLI, PAUL, Homogeneous equilibrium magmatic melts, *Faraday Soc. Trans.*, vol. 20, pp. 428-441, 1925.

² Cambridge Philos. Soc. Trans., vol. 9, No. 2, p. 8, 1850.

³ Philos. Mag., vol. 50, p. 324, 1900.

zation, and the olivine crystals that first formed settled to the bottom, the melt left would be a mixture having the composition of pyroxene and quartz.

If we now combine the two series just mentioned and consider a melt of plagioclase and pyroxene, it is easy to see that there might separate a bottom layer containing olivine and basic feldspar and a top layer containing quartz and silicic feldspar. Probably an intermediate rock would lie between.

This is our best laboratory evidence as to *how we get our series of common rocks from basaltic magma*. Even alkalic rocks might form by such a separation of crystals from a magma; leucite can be formed from orthoclase melts, leaving silica.

Filter Pressing.—A magma with 50 to 80 per cent crystals, even if in equilibrium, has different composition in the solid and liquid parts, because of the reactions shown above. If such a mushy mass is deformed by earth movements, the liquid may be separated from the crystals just as definitely as by sinking crystals.

Zoning of Crystals.—The final example of an interrupted reaction is that of fairly rapid cooling during crystallization. The early crystal of an isomorphous series may not remain long enough in contact with the evolving magma to react with it. The further growth of the crystal is then of different composition from the nucleus—the crystals are zoned. The early crystals of a discontinuous reaction series are not zoned but may be partly separated from the melt by a coating of the reaction product (pages 236 to 239).

Crystallization Stages in the Evolution of Magma.—It should be emphasized that the series of rocks derived by the separation of crystals whether by settling or filter pressing is not a series with compositions that represent the composition of any whole magma; such rocks represent only the accumulated crystals.

On the other hand, the liquid portion of a magma from which crystals settle in succession evolves through a series of compositions, and if at any time in its evolution, some portion of the magma is chilled, the rock formed will give an approximation to the composition of the liquid magma. In support of this idea it is found that the glassy rocks, which are always formed from liquid magma, show a limited range of composition as compared with the granitoid rocks, which include monomineralic rocks, formed by the accumulation and reaction of crystals.

Attempts have been made to study the progress of such an evolution of residual magma in the light of a theory that most of the minerals are isomorphic series. Anorthite grades toward albite, orthoclase, and quartz; diopside toward hedenbergite; enstatite toward hypersthene; and olivine toward fayalite. The whole idea, however, seems to ignore all other processes of differentiation, of which there may be several besides crystallization (pages 249 to 252).

Order of Crystallization.—The equilibrium diagrams give a theoretical basis for estimating what minerals should form first from certain melts. The matter is complicated by three or more features, however—by the reactions after crystallization, by the fact that magmas are more complex than the laboratory melts, and especially by the effects of mineralizers.

A study of eutectic crystallization shows little reason to expect the “normal” or Rosenbusch order of crystallization (page 44). That observed order, however, may possibly be reconciled to the reaction series as outlined. It is noteworthy that some minerals may crystallize early in normal sequence and again later as minerals “released” by reaction. Magnetite is a good example of such a released mineral, forming early in gabbro and again late by reaction of iron-bearing silicates and hydrous magma.

The order of crystallization may be changed by supercooling (page 231). A mineral that normally crystallizes early may remain liquid in a supercooled melt until some other mineral crystallizes. Thus a magma that is stirred or actively moving so that supercooling is prevented may yield an early accumulation of crystals, or even a segregation, that is very different from the early accumulation in a quiet body of the same sort of magma.¹

Textures and the Equilibrium Diagrams.—The discussion has now covered most of the factors that influence texture, the chief of which are rate of cooling and composition. Glasses result from a quick chill, and spherulites form in them when magma is slightly supercooled. The viscosity of highly siliceous melts is a factor in determining texture. Porphyries might be attributed to an early growth of a mineral in excess of eutectic proportions, the groundmass being the eutectic. But in the more common magmas some minerals grow long before the later intergrowth, phenocrysts growing large before viscosity increases enough to restrict diffusion. Other porphyries probably have had an interruption in their crystallization, coarse crystals forming at depth, and a groundmass forming after eruption to surroundings where cooling is rapid. Still other porphyries may have phenocrysts developed by inoculation or stirring of a supersaturated magma. Corroded crystals indicate a change of conditions; but the early notion that the magma must have been reheated is now largely abandoned in favor of some resorption by reaction during continued cooling.

Slow cooling of a mixed magma develops only an average granitoid texture, but two factors may make the grain extra coarse—a predominance of one mineral or a reduction of viscosity by mineralizers. Minor variations of the granitoid fabric, such as monzonitic and diabasic textures, probably depend on the proportion of constituents; diabase, for example, is common only in rocks with abundant plagioclase. Flow

¹ВОСНЕМТ, Н., Reaction rims . . . , *Zeit. prakt. Geologie*, Vol. 38, p. 177, 1930.

structures indicate movement during crystallization, and are most characteristic of surface flows.

The graphic intergrowths in rocks are puzzles. Graphic intergrowths of the same patterns are known in eutectic alloys, but it is doubtful if their occurrence in rocks indicates any eutectic, for it is known that in ternary magmas binary graphic growths vary and are not last to form. Graphic intergrowths are known also to result from unmixing of solid solutions, but although unmixing is probable, it is not proved to be the cause of any feature in rocks. Perthitic intergrowths may result from unmixing, but they differ considerably from graphic patterns. In both rocks and ore minerals graphic textures seem to be very commonly a result of *alteration*, or *late magmatic reaction*—almost a sign that reaction has occurred. Mineralizers no doubt play a part. In some phenocrysts the graphic texture may be simple simultaneous crystallization. (See the following suggested criteria and the list of further readings.)

The term graphic is not strictly defined, but clearly it should be reserved for those patterns that resemble cuneiform characters. The lattice structure, illustrated by magnetite-ilmenite, and perthitic structures seem clearly distinguishable from the graphic, but there may be gradations.

CRITERIA SUGGESTING THE ORIGIN OF MINERAL INTERGROWTHS IN ROCKS

Definition.—Intergrowth of two minerals in which the several areas of each mineral are uniformly oriented. (Coronas and poikilitic textures may lack orientation.) Common intergrowths in rocks are quartz-feldspar, perthite, feldspar-mafic silicates, magnetite-ilmenite.

1. *Binary eutectics*, probably rare in rocks. In alloys two phases may crystallize at constant temperature with a variety of intergrowths.

2. *Unmixing of solid solutions*, probably the explanation of perthites, magnetite-ilmenite, and several sulphide intergrowths. Many produce "lattice" structures; needles follow planes of greatest atomic population. The mineral present in minor amounts may be progressively excluded from the host mineral, resulting in various proportions of the included mineral. If the unmixing produces a eutectoid, the probability of graphic intergrowth is greater, but examples can not be cited.

3. *Contemporaneous crystallization*, but not a eutectic. Supposed to explain many graphic intergrowths in rocks such as quartz-feldspar phenocrysts in a porphyry; and intergrowths from a hydrous magma if water does not enter the intergrowth. The proportions of the two minerals intergrown might not be so uniform as in eutectic and eutectoid intergrowths.

4. *Recrystallization*, especially in contact zones, has produced graphic intergrowths of quartz and silicic feldspars in positions where there seems to be no chance that magma penetrated and no sign that added material replaced the original. The texture exactly resembles those produced by replacement and by simultaneous growth, but its position is different. It has been seen, for example, in place of the clay matrix in a metamorphosed shaly sandstone.

5. *Replacement*.—The intergrowth encroaches on and transgresses early structures, grains, bedding, etc., until replacement is complete. (a) Deuteric action gives myrmekite and other igneous intergrowths. (b) Possibly pneumatolytic (?) contact-metamorphic replacements. (c) Pegmatitic, chiefly quartz, with feldspar, or tourmaline, or mica or other silicates. (d) Hot and cold water replacements; much of

the quartz-feldspar intergrowth in pegmatite and especially in contact rocks is apparently replacement. There may be a gradation from deuteric and pegmatitic to contact replacements. Probably the only graphic structures formed by replacements in cold-water action are the sulphides found in secondary enrichment, and some in the salt deposits from sea water.

Deuteric intergrowths have characteristic "wormy" inclusions, in contrast with the regular "cuneiform" inclusions of some primary intergrowths. Contact replacements, however, may show "cuneiform" patterns.

EXPLANATIONS FOR ROCK SERIES

BROAD PROVINCES OF THE EARTH AS A SERIES

The characteristic rock series, both in a broad way and in some examples, have been described in an earlier section (pages 130 to 138). So far as there are great provinces, they are probably attributable to original differences in the planetesimal material that fell on different parts of the earth. A province in which rocks contain more than average titanium, like that in eastern Spain, is hard to explain except on the theory of original differences in the earth. The partial or complete fusion of the outer portion of the earth might not completely obliterate such original differences, and it must be admitted that broad provinces do not as a rule exhibit any striking extremes in composition. The differences between the rocks of continents and those of ocean basins are slight as compared with the diversity that may result from other causes in a single mass of magma. Emphasis is therefore given in the following discussion to processes causing diversity in a magma more or less homogeneous to start with.

Harker and more recently Burri and Niggli have suggested a certain amount of control of the nature of rock series by structural or tectonic features (page 130). The processes by which these act are not stated in detail, but the field occurrences support the idea of connection. Since the Triassic, the calci-alkalic group of rocks has developed chiefly in the belts of active orogeny, the earth's "hinge zone" around the Pacific Ocean and from the West Indies east to the East Indies. Potassic rocks locally border this zone; and sodic rocks are found in the broader areas beyond (see map and page 131).

EXPLANATIONS FOR ROCK SERIES IN LOCAL UNITS

The processes involved in the development of different series may differ considerably, so that it is well to review the several hypotheses and note those that are favored and those of doubtful value.

1. Generation and Mixing.—Some mixing of magmas from two sources may develop a series, but the process is probably rare. Magmas, as generated, are almost certainly different in different regions, whether from differences in planetesimal material, or in depths of supply or in completeness of fusion.

2. Assimilation.—The assimilation of a wall having a composition different from that of the magma produces a series of rocks of local distribution. It is doubtful whether such assimilation ever develops large bodies, and it is estimated that magma is rarely changed in composition by the assimilation of as much as 10 per cent of its weight of rock.

3. Differentiation.—The final and favored theories explaining rock series are theories of differentiation.

Differentiation is a process by which a magma more or less homogeneous to start with separates to form a series of rocks of different compositions. (This separation is the feature that excludes direct assimilation, though assimilated material may *activate* differentiation.)

A. The Soret action.—If different parts of a liquid solution are held at different temperatures until diffusion establishes an equilibrium, there will be slight differences in composition at the two extremes of temperature. The effect is not of a kind or of a magnitude that will suffice to explain such rock series as are commonly noted in the field.

B. Gravitative effects.—The magma in the lower part of the large magma chamber is under greater pressure than that near the top by reason of the height of magma resting on it. The difference in pressure may cause differentiation to an extent about equal to the Soret action; it is not important.

C. The streaming of gases may be a somewhat different pressure effect. Physical chemists find good reason to believe that a streaming of gases in a magma to regions of reduced pressure occurs, the gases bearing with them constituents for which they have special affinity. Possibly there is a variation in osmotic pressure related to total pressure (page 218). The magma thus becomes heterogeneous, and if it then solidifies it gives a rock series. See also the following paragraph. On the other hand, if mineralizing gases escape early from the borders of a liquid magma (as outlined on pages 213 to 216), the border magma, depleted in gases by emanation, might receive additions from its central mass by diffusion and convection. This also may well be described as a "streaming."

D. The emanation of gases from a magma may permit the formation of two kinds of igneous rock from a single original magma. If pegmatitic emanations are gaseous (page 220), they are an excellent example of differentiation by the boiling out of vapors from a liquid. Gas bubbles that separate from a magma at considerable depth may rise and react with the upper parts of the same magma, and a considerable transfer of material may result. Gas tension may also produce *lit-par-lit* injections and ribbon injections. A final suggestion is that emanation from a deep layer or zone of magma may react on the rock already solidified above and produce a sort of differentiation in it by introduction.¹

¹ GILLSON, J. L., Petrography of the Pioche district, Nevada, U. S. Geol. Survey Prof. Paper 158 D, 1929.

E. The separation of partly miscible liquids probably explains the occurrence of some sulphides of magmatic origin. It has been shown that liquid silicates also may be immiscible. The field evidence is strong but not entirely conclusive for such immiscibility in magmas. The laboratory evidence indicates little chance of it being important, but the difficulty of approaching natural conditions as to pressure and water content in laboratory melts leaves the question still open (see pages 222 to 224).

Igneous differentiation may possibly occur (especially in *lit-par-lit* injection and through porous wall rocks) by some process of separation of liquids analogous to the separation of oils into fractions as they pass through clays. Magmas, however, are electrolytes and may not be adsorbed as oils are. There is therefore some doubt as to the process.

F. Differentiation during crystallization is the process favored as explaining logically many of the rock series found in nature. Differentiation may result from (1) a localization of the formation of crystals or (2) a separation of crystals from the residual liquid in which they are suspended.

1a. It has been supposed that the cooling borders of a magma reach the temperature of crystallization before the center, and as crystals grow the liquid around them is kept uniform in composition with the central portion by diffusion. Diffusion is certainly too slow in viscous silicate melts to permit much differentiation by this method (see page 210).

b. Crystals may grow near the cooling borders of a magma and continue to grow by slight diffusion, if a general circulation of the magma brings into contact with the crystals a continuous fresh supply of the molecules needed for their growth. This is the probable explanation of those exposures in which the border grains are coarse. It is about the only possible explanation for zoned border phases in which heavy magnetite or olivine rock is formed both above and below lighter rocks, such as that at Raana, Norway, or that at Carrock Fell (see page 133 to 136). Convection is also the only satisfactory explanation of banded structures in the differentiates, especially if the bands formed in vertical position along the walls of a large intrusion and show a flow structure parallel to the banding.

c. A third process of crystal growth is more obscure but none the less real. Assimilation whether near the surface or at moderate depths seems to produce some rock series which include other factors than simple contamination. The rocks are not the results of simple syntaxis but of syntectic differentiation. It is suggested that the assimilated or contaminating materials may act almost as a catalyst speeding up some process of differentiation. Many included fragments of wall rock in magmas seem to have caused a crystallization of some material from the magma that would not otherwise have formed. The minerals segregated (page 132) were magmatic, for there is no other adequate source of such

material, but the process is evidently controlled in some way by the inclusion.

These known examples of syntectic differentiation are all minor features of some large mass, but it has been suggested that much larger differentiates are results of a similar syntexis. The arguments against important assimilation (page 228) are also against the importance of syntectic differentiation. The larger rock series, especially the alkalic rocks, which have been attributed to such a process, can be explained in other ways.

2a. Most crystals that form early in a magma are heavy enough to settle. Olivine will settle in a crucible melt. In nature crystals that form near the cooling wall and roof are likely to settle into the deeper and central parts of the chamber, where the hotter magma will redissolve them, creating a certain amount of heterogeneity in the liquid and ultimately a rock series if the heterogeneous magma solidifies. If the process continues, however, a stage may be reached at which crystals settle through to the bottom of the magma chamber and accumulate as a layer of mushy half-crystalline magma, later to form rock. The first crystals to settle in this way are not the average of the rock, and so the rock formed at the base will differ from that above.

It is difficult to imagine crystals around a millimeter in diameter settling many miles to reach the bottom of a very large body of cooling magma. Such a notion of differentiation is an extreme extrapolation of a process seen in experiments, in which the crucible used was perhaps 2 cm. deep. Although there need be no doubt of the tendency to settle, the crystallization produces other effects, which overshadow the settling in importance. The crystals grow as a crop of considerable width near the cooling surface. The magma with crystals in it through this zone is heavier than magma away from the wall, where it is hotter and free from crystals. The whole zone of magma containing crystals will start into circulation (pages 196 and 197), and the crystals will go down with the magma. This convection will be much more rapid than the crystal settling. When the magma in convection flows inward along a floor, any crystals in it may settle a few inches or a few feet and thus be removed from the general circulation. The banded and fluxion structures associated with large intrusives are strong evidence of this modification of the simple settling process.

It is not here suggested that convection is a process making magma differentiate, but it is a process that modifies the simplicity of crystal settling and offers logical explanations of the distances through which crystals travel and of the structures in differentiated rocks.

b. *Filler Pressing*.—A partly crystalline magma may be deformed by earth movements. Such a process might drain off some liquid from the aggregate of crystals, and as the crystals and liquid differ in composition

this process would result in different rocks. This is the common notion, and probably a correct one, for the formation of pegmatites and ores as emanations from igneous rocks; but the pegmatites form in some other ways also.

Just as the gaseous emanations from the deep parts of a chamber may be introduced into the earlier upper rocks of the chamber, so this late, residual, filter-pressed magma may modify some of the earlier rocks by a sort of introduction, hardly to be distinguished from deuteric reaction.

A suggestion has been made that deformation or intrusive movements in a partly crystalline magma might cause segregation of early crystals into "flow layers" by some sort of interference of crystals as they moved, but the idea lacks experimental support.

PROBABILITY OF COMPLEX ORIGIN OF SERIES

With all these various possible modes for the development of rock series, it is not necessary to select just one of them as a favored theory.

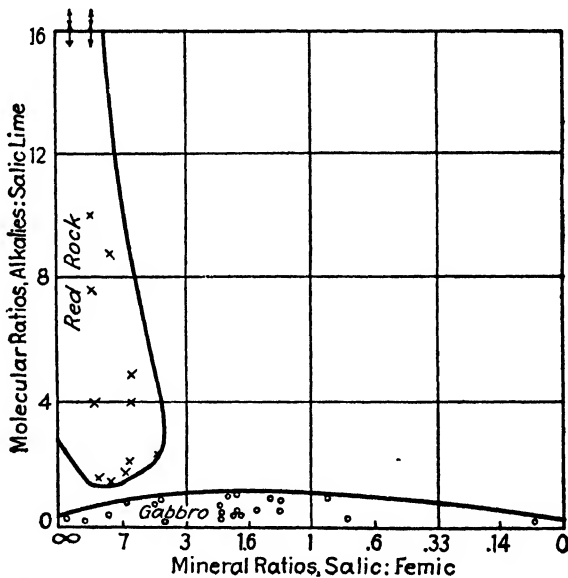


FIG. 157.—Diagram of the rock series in the Duluth gabbro formation. The variation in the gabbro is in a different direction from that in the red rock though both series developed in the same chamber. The gabbro ranges from salic to femic but has no alkalic phase. The red rock ranges from alkalic to calcic, but has no femic phase.

Several processes probably operate in the consolidation of each rock mass and the evidence of the complexity may be summarized as follows:

We are certain from experimental work that (a) melts assimilate, (b) volatile substances migrate and escape, (c) crystals and separated molten sulphides settle, and that each of these processes would make characteristic differences in the magma.

Some rock masses exhibit two kinds of variation in a single chamber (Fig. 157).

The trends of different series are in different directions (Fig. 111), and the curves in some series are linear but in other series they are broken.

Certain groups of rocks such as the alkalic masses are of three or more distinguishable kinds, and it is unlikely that all formed from normal magmas by the same process.

The gravitative adjustments, which seem so conclusively shown in some differentiated sills, can not explain masses like that at Raana, Norway, in which heavy differentiates occur both above and below the light rocks (pages 133 and 134).

CONDITIONS FAVORING DIFFERENTIATION

Large volume and the consequent slow cooling are favorable to differentiation processes. Few sills less than 500 feet thick grade from gabbro

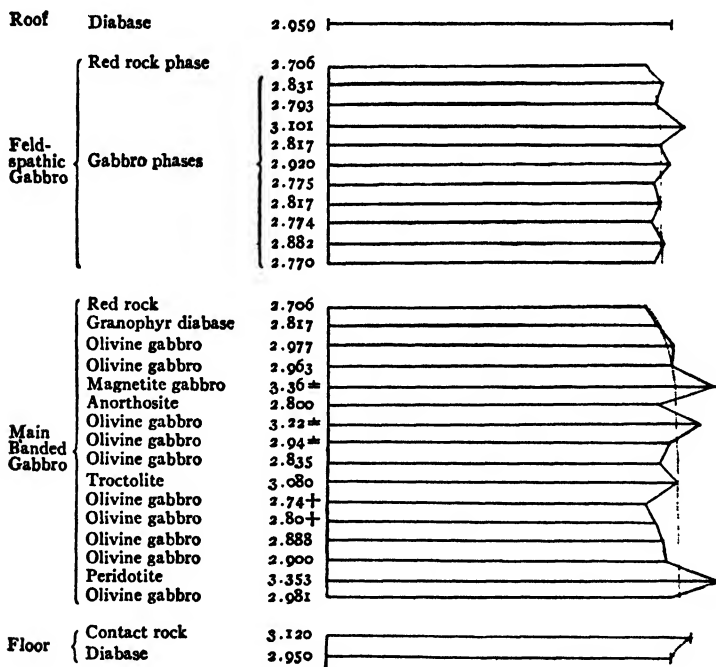


FIG. 158.—Specific gravities of a series of Duluth rocks in "stratigraphic" order.

to granite as the thicker sills do. A hot wall rock would result in slow cooling of even small masses and favor differentiation.

Certain features of magma composition are favorable: (a) Mineralizers probably increase fluidity; pegmatites are more differentiated than less hydrous dikes; (b) basaltic magma is more fluid than rhyolite magma and probably allows more rapid settling.

There is also a greater chance for differentiation in basaltic magma because of the high specific gravity of the early crystals, than in rhyolite where the dominant minerals are all light.

Orogenic movement may cause differentiation by filter pressing, where no differentiation would otherwise occur.

CRYSTALLIZATION DIFFERENTIATION AND SPECIAL CASES

The recent popularity of the theory of differentiation by crystal separation is probably a result of its adaptability to special cases. The following notes may serve to show its wide range of applicability.

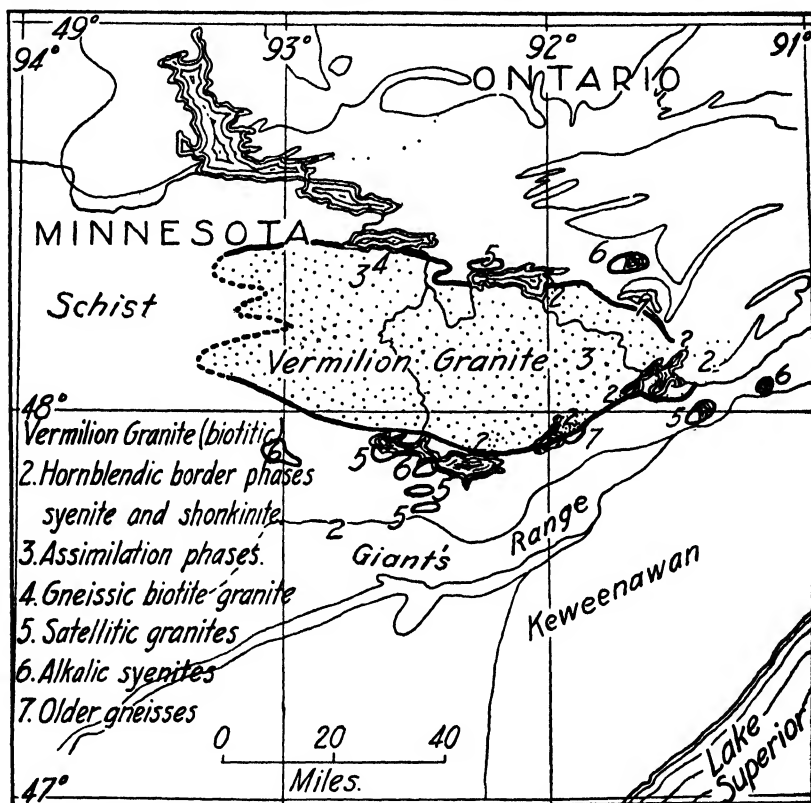


FIG. 159.—The Vermilion batholith of Minnesota and Ontario and its surroundings. The border phases are hornblende around a biotite granite. The distribution of the small stocks around the batholith suggests that they are related to the main mass, but several of them are more alkalic than the batholith.

Volcanoes pour out lavas of different compositions at different times, because the magma in the reservoir below is differentiating. There may even be an alternation by eruption. When the upper, siliceous layer has been extruded, the next lower layer may contain crystals and be more basic but still be eruptible. After the eruption of the second layer the

reservoir may again develop an upper, siliceous layer for the next eruption. A possible complication is that the reservoir itself may be a satellite and receive at times an addition of primary magma from depth.

The *gravitative arrangement of differentiates* is the basis of the crystal-settling theory, and that theory explains fairly well the occurrences known. The gravitative order, however, is not perfect, and it is really not to be expected, because the heaviest minerals are not always the first to crystallize (Fig. 158). Note also that some series in gravitative arrangement may be enclosed in a border that was chilled before it had time to differentiate.

The *alternation of banded differentiates* is best attributed to crystal settling during convection with rhythmic effects due to cooling and injection.

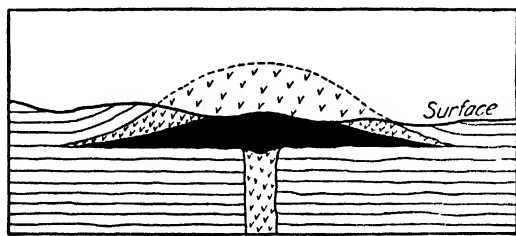


FIG. 160.—Diagram of a possible method of segregation and erosion, to produce an area of igneous rocks with a center more basic than the outer margin.

Basic borders around large intrusives are explained as early phases, solidified before differentiation had time to change the magma as far as the central part of the mass was changed. The common assumption, however, that batholith magmas evolved from basaltic to granitic composition and that the borders chilled early, producing more-basic rocks than the center, needs a check by careful work. Most cases may be explained by a border accumulation of the early crystals in the granite magma without any appeal to early basaltic magma. The probability of an early basic magma is strong in the Vermilion batholith (Fig. 159), which has not only a border phase but also some related basic dikes, which could hardly form by a segregation of crystals.¹

The rare case of a *siliceous border* around a more-basic intrusive, illustrated by Haystack Mountain, Mont., is not so readily explained. The Snowbank stock in Minnesota (Fig. 134) has a granite border resulting from later injection along the sides of an early plug. Again, a basic center may result from erosion of a lenticular mass like a laccolith, which might form a thicker basic differentiate in the center than at the edges (Fig. 160).

¹ GROUT, F. F., Geology and magnetite deposits of northern St. Louis County, Minnesota, Minnesota Geol. Survey Bull. 21, p. 38 and plate 6 A, 1926.

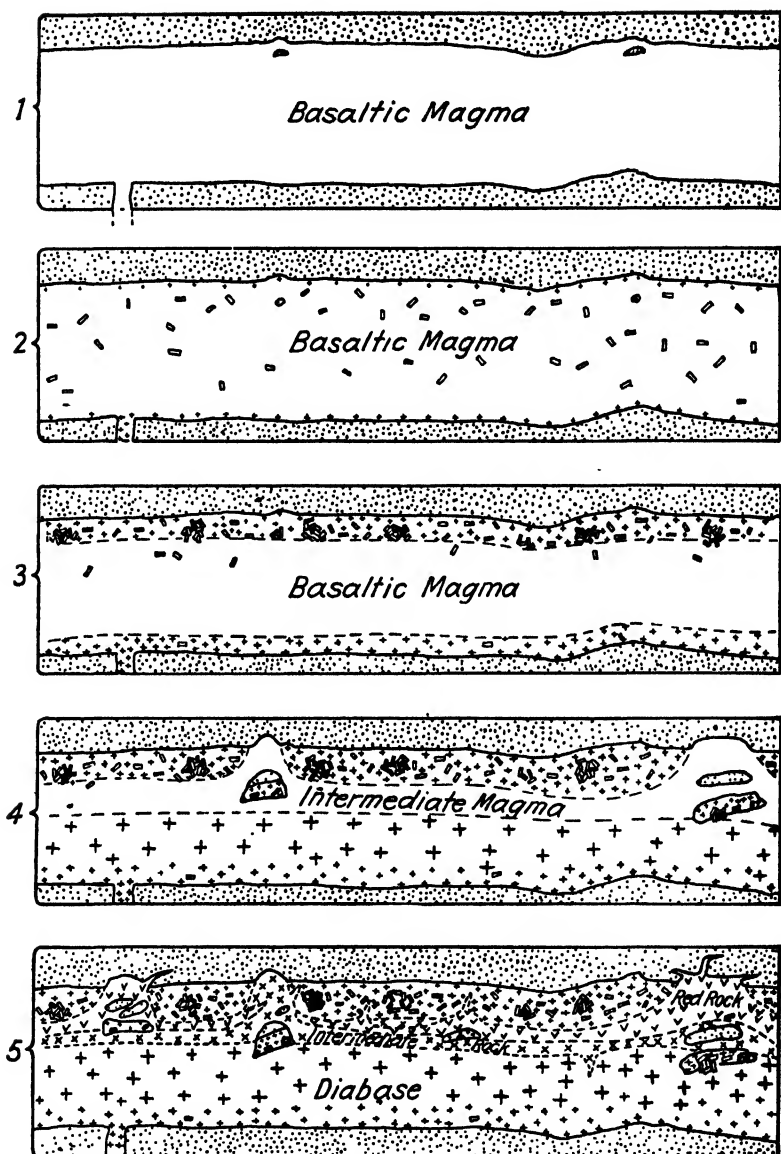


FIG. 161.—A series of diagrams based on field observations at Pigeon Point, Minnesota. This is believed to indicate the succession of events when a magma differentiates into a series including both anorthosite and granite. 1. The main sill chamber is filled. 2. Phenocrysts of plagioclase rise. 3. Segregation of anorthosite and cooling of the borders of the sill. 4. The main diabase solidifies in the lower half of the chamber leaving the magma more alkaline and siliceous. Some disturbances. 5. The final differentiation leaves a granitic residual magma that forms red rock, after some further mechanical adjustments.

The fairly common case of random distribution of different rocks in a batholith is probably a result of a movement of some differentiates after others had solidified.

Complementary dikes near an intrusive are not usually contemporaneous. Many basic dikes represent offshoots at an early stage of the development of the main magma, and the later dikes were formed from aplitic residual liquors from the nearly solid rock. Some few cases where two magmas seem to have been fluid at the same time and yet to have been derived from a common source may be explained by the melting of settled crystals, producing a heterogeneity in the fluid.

Petrographic provinces are perhaps best illustrated by the change in the rocks at the Rocky Mountain front in Montana. Alkalic rocks occur in the scattered mountain groups east of that front and sub-alkalic rocks west of it. They may be derivatives of the same magma, eroded to different relative depths or brought into place by different tectonic forces. The alkalic differentiates have a distinct tendency to occur marginally in satellitic bodies at high horizons.

The anorthosites are probably floated up (Fig. 161). They could hardly have been fluid, because they have a high temperature of fusion, and no extrusives and few dikes have such compositions.¹

The frequency distribution of rocks (Fig. 103) is explained by the contrast in (1) bodies with floor, clustered about basalt and (2) those deep masses with no known floor, which have differentiated and have only the tops exposed, these being clustered around granite.

Nephelite rocks may be formed by some disturbance of equilibrium, producing albite and nephelite (and water) from a magma that would normally have an analcite base. Or, again, the reaction giving leucite and quartz from orthoclase (Fig. 150) may be followed by a reaction of leucite with albite magma, forming orthoclase and nephelite.

There is still much discussion as to the origin of alkalic rocks and of anorthosites. See the lists of further readings. It is common in such discussions to note that many magmatic reactions are reversible, and that equilibria are easily disturbed. The disturbance may arise from temperature, pressure or concentration, but the favorite appeal is to a disturbance caused by a change in mineralizer content. If some such unknown feature as an exceptional water content modifies the early stages of crystallization, the later stages may be so greatly changed that the rock will seem to be derived from a magma differing in many other ways (pages 242 to 244).

PROBLEMS IN DIFFERENTIATION

On the other hand, there are some rock series that appear to be a result of differentiation but are not easily explained by unmixing of sulphides, nor by volatilization, nor by crystallization and processes of settling or filtration. One such case is that of an intrusive with a heavy

¹ But Lodochnikow (Jour. Geology, vol. 33, p. 153, 1925) attacks the idea.

mineral such as olivine or ilmenite segregated at both the top and bottom of a mass that has only lighter minerals in the center (pages 133 to 136). Not many such series have been well studied. As a second problem, there are masses showing two kinds of series, one easily explained by crystal settling but the other so different that it suggests a very different sort of process (page 252). Third, the settling of a mineral should normally be followed by the settling of a different mineral giving a change in the trend of differentiation, but some series show such a straight continuous series as to suggest that they are not a result of crystal separation.

In strict analogy with the alkali concentration from plagioclase melts there should also be an accompanying iron concentration from pyroxene melts during crystallization.¹ If such a concentration of iron occurs, it is not at all certain that the fractional crystallization of basaltic magma will cause the residual liquid to approach granitic compositions.

No theory yet proposed gives a really satisfactory explanation for the late dikes of two kinds, silicic and basic, related to an intermediate stock, nor for segregations around inclusions and orbicules, nor for the sharp separation of silicic and basic differentiates in a chamber where intermediate rocks are scarce, nor even for the many alternations of rhyolite and basalt in a series of lava flows.

There is also a great deal of disagreement as to the manner in which volatile constituents of magmas migrate within the magma and emanate from it, as they so evidently have done.

RECENT CONTRIBUTIONS AND SUMMARY

A summary of 50 years' progress in petrography has been compiled by Miss Bascom.² Probably the most outstanding work is Dr. Bowen's emphasis on crystallization as the sole control in the evolution of magmas, and Dr. Fenner's clear demonstration that at Katmai and other places some other explanations are needed.

Besides the favored modern theories of differentiation by crystal separation it is wise to retain as working hypotheses the migration and escape of volatiles, the mingling of magmas, the processes of assimilation, and the separation of partly miscible magmas. These may be modified by other magmatic processes, some recognized but probably others unknown. With puzzling differentiates, however, the student should be on guard against promptly referring all the results to some unknown process occurring "below." The exposed rocks may show what to expect below. See the list of further readings.

¹ FENNER, C. N., The crystallization of basalts, *Am. Jour. Sci.*, 5th ser., vol. 18, p. 229, 1929; but see also R. J. LUND, Differentiation in the Cape Spencer flow, *Am. Mineralogist*, vol. 15, p. 563, 1930. The iron does not concentrate far.

² BASCOM, FLORENCE, Johns Hopkins Univ. Studies in Geology 8, pp. 33-82, 1926, Baltimore.

CRITERIA TO DISTINGUISH DIFFERENTIATED MASSES FROM COMPOSITE INTRUSIVES
AND THOSE THAT VARY BY ASSIMILATION1. **Separate intrusions.**

Intrusions with sharp contact (if not faulted).

Apophyses of one rock in another.

Rapid gradation outward or downward to more-silicic rocks.

Xenoliths and xenocrysts of earlier rocks in later.

Difference in degree of metamorphism of different intrusives.

2. **Differentiation.**

Gradational contacts unless disturbed at a late stage.

Silicic rocks in the central to upper parts.

The xenoliths of early phases seem related but are not commonly abundant.

3. **Assimilation.**

Gradational contacts.

Composition grades toward country rock, usually calcic, silicic or aluminous.

In a few cases, also, zoned around xenoliths.

Odd minerals in an igneous rock—sillimanite, staurolite, cordierite, garnet, graphite.

Embayed in certain rocks more than others.

PART IV

THE PETROGRAPHY OF SEDIMENTARY ROCKS

GENERAL FEATURES

Sediments are deposits of solid material (or material in transportation that may be deposited as solid) formed on the surface or in the outer crust of the earth under normal temperatures.



FIG. 162.—Nearly horizontal bedding in sandstone, at Sandstone, Minn. *(A photograph by Sandstone Quarries Company.)*

Most sedimentary rocks are bedded, technically called stratified, deposits. Beds or strata differ from one another in texture or composition or both (Figs. 162, 163, and 173). The thinner divisions within a bed of considerable thickness are called laminae. Many of them differ from each other only in color, though others may differ in texture and composition also. Laminae are rarely more than half an inch thick (Fig. 164). Thicker bands by general usage are called beds or strata, and they are commonly separated from the beds above and below them by bedding

planes along which there is little cohesion. A group of beds of similar material constitutes a formation. These several structures may be very smooth and uniform over great distances, as in some clays and chemical and organic sediments, but in sandstones and mechanical sediments the beds are likely to be very irregular. Practically all beds ultimately



FIG. 163.—Bedded sediments folded in a gentle anticline. The Key, east of Georgian Bay, Ontario.

grade into other materials laterally, or pinch out so that in a large or small way they are lens shaped.

Most sediments are fragmental but more or less cemented by minerals deposited between the fragments by reactions in circulating water solutions. These solutions may bring soluble minerals from a distance or locally recrystallize those in the fragments.

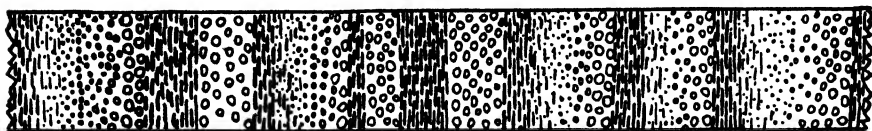


FIG. 164.—Diagram of the textural variation in a series of beds of sandy clay. The normal gradation is from coarse grain at the bottom of a bed to fine grain at the top. The next bed begins abruptly with coarse material over the fine grained top of the first bed. This shows the original position of tilted beds, tops to the left.

The rocks formed by organisms and those formed by chemical action in water solution are usually included with sediments. The group might well be designated "aqueous and eolian," but by custom the term sedimentary has been extended to cover all.

Fragmental rocks resulting from leaching and from weathering are also included, though the particles may never have moved so far as those we think of normally as sediments. Only those petrographers who make metamorphism a synonym of alteration would class weathered rocks as metamorphic.

The division between sedimentary and metamorphic rocks is arbitrary. Processes of change may begin in sediments as soon as they are deposited. If the change is great, and especially if it has taken place under high pressure, we call the rock metamorphic, but if slight we still class the rock as sedimentary. Certain changes that occur before the succeeding beds of sediments (more or less uncertain as to volume) are deposited, are classed as primary changes or processes of diagenesis, and are included in sedimentary processes. These include concretionary effects, cementation, and even a certain amount of dehydration (clays), recrystallization (limestone) and replacement (dolomite). By definition diagenesis denotes physical or chemical changes in sediments before consolidation, while they are still influenced by the conditions of deposition. The term consolidation, however, as applied to sediments is very indefinite.

RELATIVE ABUNDANCE OF SEDIMENTS

Nearly three-fourths of the rocks exposed at the surface are sediments. A preliminary estimate as to the relative abundance of the three chief clans follows.¹

The chief sediments	In measured geologic col- umns, per cent	Calculations from average rock, per cent		
Clays and shales.....	46	80	82	87
Sandstones and other mechanical sediments...	32	15	12	8
Limestones.....	21	5	6	5
Others.....	1			

It is evident that the study of sediments should be largely devoted to three clans and that detailed systematic names for the other sediments are hardly justified.

METHODS FOR SEDIMENTARY PETROGRAPHY

The description of sediments presents a problem that differs in several respects from the one involved in the description of igneous rocks. The minerals normally associated are different; there is a much greater tendency for a concentration of one mineral to the exclusion of others;

¹ LEITH and MEAD, "Metamorphic Geology," pp. 59-68, 1915. F. W. CLARKE, "Data of Geochemistry," 5th ed., U. S. Geol. Survey Bull. 770, p. 34, 1924. W. J. MEAD, The Average Igneous Rock, Jour. Geology, vol. 22, p. 772, 1914.

there are many more fine-grained, even colloidal materials; and many more specimens prove to be so friable that thin sections can be made with difficulty if at all. Finally, the variability of a formation laterally may be considerable, but the succession of beds vertically shows such extreme changes that it is difficult to determine how to sample a formation and how much to include in a description. Field methods are necessarily adjusted to the extent of the laboratory work that it is planned to undertake.

Field Methods.—Describe first the largest units recognized, then those of the next order and so on down to the smallest. For each main station or formational unit cover the following particulars in more or less detail. For the greatest detail it is desirable to have a memorandum sheet in the notebook covering minor points to look for, but the list here given can be memorized and used anywhere.

1. Location.
2. Name of rock.
3. Size and shape and topographic development.
4. Bedding features and large structures.
5. Color¹ when fresh.
6. Luster.
7. Composition; size and shape of grain and minerals.
8. Hardness under a hammer.¹
9. Fossils.
10. Concretions.
11. Variation.
12. Weathering
13. Specimens collected (Problem 1).

Laboratory Methods.—The naming and description of sedimentary rocks are ordinarily field problems of no great complexity. As the rocks may show rare peculiarities and as the need for more accurate work occasionally arises there are some fairly well known more detailed methods. Probably most of the careful descriptions of sediments in recent years have been made in the effort to correlate formations encountered in drilling for oil. If the cuttings are washed and inspected, tested with HCl and again inspected (possibly with a pocket lens or binocular microscope), the formations are satisfactorily identified in probably 95 per cent of the holes drilled in an area close to some earlier drilling. If such simple treatment at the drill-rig is insufficient for the more important correlations the work progresses to more and more difficult operations until the whole list of methods may be applied. Fossils, especially microfossils, are of particular value in correlation, but their recognition requires another course of study. The petrographic methods here given have nevertheless a considerable value. The fossils may be entirely

¹ For quantitative methods on color and hardness, see Readings on Methods.

absent from some formations, but petrographic features are never absent. It is found also that the minerals and textures serve to subdivide some considerable formations that may be constant in paleontologic characters.

The following outline of procedure is given as if it were intended to carry out the work in the greatest detail. In practical work, of course, the procedure is interrupted as soon as the data available are sufficient for the purpose in mind—say for correlation.

Outline of Procedure.—This is applicable to the common sediments, shale, sandstone and limestone. See also Milner's "Sedimentary Petrography."

1. Quarter the sample to the desired size.

2. Weigh out a portion. Use smaller amounts if there are fine uniform sands; amounts up to 50 grams if the grain is 1 mm. or if the material is mostly clay or limestone. Weigh a separate sample, commonly 10 grams, into another dish to determine the loss in acid.

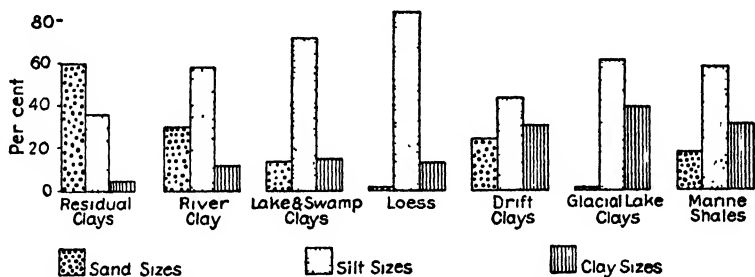


FIG. 165.—Diagrams of the "grade-sizes" of a series of Minnesota clays. This is a small form of "pyramid plot" or histogram. For a group of sediments not restricted to clays, more columns are used and the limiting sizes are stated, as in the paper by Wentworth in Iowa University Studies, vol. 11, No. 11, 1926.

3. Treat both with dilute HCl and boil to loosen the grains (see Problems 3 and 4, pages 13 to 16). Record effervescence.

4. Filter the 10-gram sample, wash, dry, and weigh; calculate the loss as mostly carbonates.

5. Treat the main portion to remove the finest and separate the coarse into fractions (Problem 3).

6. Screen the dried sand if it shows a variety of sizes and such detail is desired. Fossils occur largely between $\frac{1}{4}$ - and $\frac{1}{8}$ -mm. sizes.

7. Separate each sand and silt by pure bromoform, and in some samples by other methods. The heavy residual minerals are to be expected chiefly between 0.5 and 0.05 mm. in size.

8. Weigh each separate and study for shapes of aggregates and grains, crystal habit, mineral nature, polished or frosted or coated surfaces, internal crystal structures, fossils, streaks, hardness, and toughness. Shale lumps, loosened cleavable feldspars and other cleavable minerals, or loosely adherent coatings are especially noted in sands for mortar and

concrete, as they are the cause of weakness in the product. "Dark ground illumination" or a black paper mount for the sands may be useful. The minerals may be tested in refractory oils, and permanent mounts are made in air, in balsam and in piperine or hyrax with a higher index than balsam (see also page 284).

9. Diagram the results:

- a. Range of sizes of grains (Figs. 165 and 166).
- b. Shapes of grain (and relation of shape to size).
- c. Minerals; the proportion of sand, lime and clay; the freshness, shape, and size of sandy minerals.
- d. Inclusions in the common minerals.
- e. The amount and nature of the heavy residual minerals, conveniently by a scale from 1 for rare to 10 for very abundant minerals.

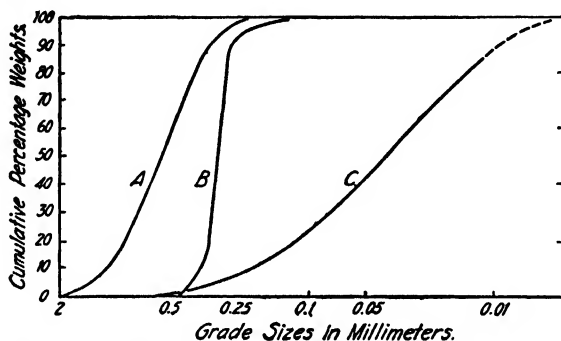


FIG. 166.—Cumulative diagrams of the grade size of three sediments. (After Holmes.) A. Red Crag sand. B. Beach sand. C. Glacial brick earth. The sizes used for points of division are arbitrarily selected and other sizes can be marked at the proper places on the scale at the bottom. This makes it possible to transform the cumulative diagram into a pyramid of any desired type. It is even possible to transform a pyramid into a cumulative curve, and back into a pyramid based on other division points.

Besides this systematic procedure (applicable especially to drill cuttings but useful also in more general studies) sediments may be tested by chemical analysis, qualitative tests and stains, and by the more ordinary optical methods on thin sections. Dense rocks such as flint can be broken by careful work with a hammer into chips even thinner than a thin section, and the chips may be mounted in oil or balsam for a study of the texture. Porosity and true and apparent specific gravity tests may be of some service. Such sediments as coal may be polished and etched for study by reflected light.

Quantitative work on minerals, especially on concentrates, may involve counting the grains of each mineral under standard conditions.¹

The applications of the methods are numerous and are noted along with further suggestions in the readings listed.

¹ The necessity of exactly standardized methods is shown by Smithson, *Geol. Mag.*, vol. 27 pp. 134-136, 1930.

CLASSIFICATION OF SEDIMENTS

It might be logical to make main divisions of aqueous and eolian rocks on the basis of origin, under three heads—precipitates, organic remains, and mechanical sediments—with subdivisions indicating where or how the deposits accumulated. This, however, would put limestones of similar general appearance into more than one class and leave about 90 per cent of our sediments in the third class.

The primary classification of sediments here used is mineralogical and serves as a basis for description. Probably 99 per cent of our sediments consist of clay, sandstone, limestone, and mixtures of these. Those sandy textured rocks in which the grains are calcite are most conveniently treated as limestones. Many minor varieties have special names, which characterize them with reasonable exactness, but the three main groups vary so widely that it is very unsatisfactory to stop with such group names. There are needed in the first place some qualifying adjectives.

Clay is an argillaceous rock.

Sandstone is an arenaceous rock.

Limestone is a calcareous rock.

Coal is a carbonaceous rock.

Iron ore is a ferruginous rock.

Salt is a saline rock.

Various other adjectives refer to less common rocks.

The terms are used mostly to describe mixtures. We seldom say "argillaceous rock" when we mean clay, but if a sandstone contains some clay, it is an argillaceous sandstone. Most of the terms are clearly mineralogical, but "arenaceous" may be textural. Usually it implies quartz as well as sand sizes of grain.

Some attempts have been made to apply a uniform system to all sediments, but none has been accepted as of general application. Several of the proposed names, however, are of value and the system may be outlined in brief.¹ The papers listed in the readings carry the discussion much farther.

1. Beginning with origin, we may divide as follows:

Primary:

(Pyrogenic = igneous, see above.)

(Atmogenic = snow, exceptional.)

Hydrogenic, mostly precipitates.

Biogenic, mostly shells and skeletons.

Clastic:

Pyroclastic = tuffs and breccias.

Anemoclastic, mostly wind erosion.

Atmoclastic, talus breccia, etc.

Hydroclastic, mainly sands, clays, etc.

(Bioclastic is artificial.)

¹ GRABAU, A. W., Classification of sedimentary rocks, *Am. Geologist*, vol. 33, p. 244, 1904; "Principles of Stratigraphy," Chaps. 6 and 17. A modified system was recently outlined by Janshi Sen, *Pan American Geologist*, vol. 41, p. 176.

In studying the origin of a sediment, however, we usually find it modified by processes of diagenesis (page 262), which are very diverse and require the addition of many qualifiers in the nomenclature.

2. Composition may be shown by the prefixes calc (i)-, silic (o)-, argill (o)-, ferro-, etc. which are almost self-explanatory. This list is not sufficient, however, to cover some special varieties. Neither does it show whether the prefix refers to the grains, or some introduced cement or both. The prefixes are satisfactory only for pure material, and a further mineralogic prefix is needed for common mixed sediments.

3. Textures are:

Rocks are:

Coarse = rudaceous
Medium = arenaceous
Fine = lutaceous

Rudites (= psephites)
Arenites (= psammites)
Lutites (= pelites)

To illustrate the use of the system consider a rock that megascopically is a medium-grained limestone. On close study it may prove to be a "biogenic hydroclastic calcarenite." This tells so much about the rock in three words that it is a most excellent system. Moreover, the terms are not so complex or obscure as to be difficult to understand. The limitations of the system, however, should be clearly understood.

The value of terms for sediments is chiefly to be judged by three points: (1) They should present a definite picture of the material; (2) they should suggest a mode of origin; and (3) they should be as simple as consistent with the necessity of covering the first two points. In the example of the preceding paragraph the term "granular limestone" fails to suggest the origin so clearly shown by the slightly longer systematic term. This is an example of an almost universal lack of definiteness in the terms used in megascopic descriptions of the abundant sediments, clay, limestone and sandstone. Each of these great groups needs a nomenclature, but possibly no single system can be adapted to all three.

In contrast with this consider the cherts. Many of these contain remains of organisms that were originally calcareous. The rock is a "biogenic calcilutite with fossils, replaced later by hydrogenic silicolutite." The term chert, however, conveys to most petrographers just as definite a picture of the material, just as definite a suggestion of origin, and is only about one-tenth as long. It would be absurd to abandon the term chert in favor of the longer, systematic term.

Again, if one heard of a "pyrogenic hydroclastic silicarenite," he might not be certain that it was an arkose that was referred to; but can any better term for arkose be evolved from the system? No doubt the term arkose is best retained.

In this class of satisfactory terms with chert and arkose are diatomaceous earth, rock gypsum, salt, lignite, cannel, travertine and others.

Terms indicating the agent of genesis and the agent breaking up a sediment mechanically are valuable, then, chiefly in connection with the broader terms for the common sediments, sandstone, limestone and clay, most of all for limestones.

Uncertainty arises with sands, because the mineral prefix silic-o is not enough to distinguish quartz from various silicates like feldspar; and even if another term were found for feldspar, there are about 200 other minerals that may locally be abundant in sands.

Again, the genetic terms for clays, the most abundant of all the sediments, are not clear. Probably most clay minerals are formed by weathering from other silicates. Are they hydrogenic? Furthermore most of the clays become fine grained during this same weathering process. Are they hydroclastic? The process of weathering is complex and may not be entirely due to hydrous agents. Certainly the residual clays should be distinguished from transported clay, but how will the difference appear in the systematic name? The answers are difficult to find.

SIZES OF GRAIN IN SEDIMENTS

Sediments are largely defined by sizes of grain, which are best standardized. Convenient sizes for petrography are shown in Table XII.¹

TABLE XII.—GRADE SIZES OF FRAGMENTAL GRAINS

Pieces		Aggregates	Cemented rock
Boulders	Roundstones.....	Boulder gravel	Boulder conglomerate
256 mm.		Cobble gravel	Cobble conglomerate
Cobbles		Pebble gravel	Pebble conglomerate
64 mm.		Granule gravel	Granule conglomerate
Pebbles		Very coarse sand	
4 mm.		Coarse sand	
Granules		Medium sand	Sandstone
2 mm.		Fine sand	
Very coarse sand grains.....		Very fine sand	
1 mm.		Silt	Silt stone
Coarse sand grains.....		Clay	(Clay stone?)
$\frac{1}{2}$ mm.			
Medium sand grains.....			
$\frac{1}{4}$ mm.			
Fine sand grains.....			
$\frac{1}{8}$ mm.			
Very fine sand grains.....			
$\frac{1}{16}$ mm.			
Silt particles			
$\frac{1}{32}$ mm.			
Clay particles including colloids.....			

A 35-mesh sieve openings $\frac{1}{2}$ mm.; a 60-mesh sieve $\frac{1}{4}$ mm.; a 120-mesh sieve $\frac{1}{8}$ mm.

Mixtures rarely have more than two of the four main groups in quantities that are essential, say more than 10 per cent. Glacial till is the exception and is known to be a mixture. Other mixtures are sandy gravel and gravelly sand, silty sand and sandy silt, clayey silt and silty clay, sandy clay and clayey sand. Uniform rounded grains have a greater porosity as a rule than mixtures in which small grains fill the spaces between coarser ones. The clay sizes will be separated in the discussion following.

CLAYS AND SHALES

Clays are commonly defined as the natural plastic earths but include some non-plastic earths which are largely composed of hydrous compounds of alumina and silica. Some petrographers would base the definition on size of grain, specifying those disperse forms of mineral grains in which particles smaller than 0.002 mm. predominate; but many commercial

¹ WENTWORTH, C. K., Jour. Geology, vol. 30, pp. 377-392, 1922.

clays have only a small proportion of such fine particles. It would be nearer average to use as the upper limit of the clay size $\frac{1}{256}$ mm., about 0.004 mm. Some of the properties of clay, especially plasticity, are related to the fineness of grain, many grains being so fine as to be classed as colloidal.

Under this definition some clays are residual from the weathering of feldspathic rocks, but most are transported and deposited by water. In fresh water the deposition of fine particles may be very slow, but in salt water the colloids are coagulated, and the sediment settles more rapidly.

Shale is typically a hard laminated clay. Some shales, however, may lack one of these two characteristics, they may be hard and not laminated, or laminated and soft. Most shales are formations that have been buried under other sediments; recent surface deposits are not called shale. The lamination is commonly detected by the ease of parting or cleavage parallel to the bedding and is supposed to be due to slight variations in the texture of successive very thin layers but may be due to some process of sedimentation, more or less obscure.

Argillite is a clay or shale hardened by crystallization, and *slate* is a similar rock with a secondary cleavage. See Metamorphic rocks.

It will be noted that the definition of clay referred to hydrous compounds of alumina and silica without giving mineral names. There are several minerals in the group that show considerable variation in the proportion of silica, alumina and water, possibly even to the extremes of hydrous alumina (gibbsite) and hydrous silica (opal); many include other constituents either combined or adsorbed. The following are recognized with some certainty in various clays. (See list of further readings.)

Clay mineral	Estimated composition	Birefringence
Kaolinite.....	$\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$	0.005(—)
Nacrite.....	$\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$	0.006(+)
Dickite.....	$\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$	0.006(+)
Halloysite.....	$\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot ?\text{H}_2\text{O}$	0
Allophane.....	$\text{Al}_2\text{O}_3 \cdot \text{SiO}_2 \cdot ?\text{H}_2\text{O}$	0
Anauxite.....	$\text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot 2\text{H}_2\text{O}$	0.006
Pyrophyllite } Agalmatolite }	$\text{Al}_2\text{O}_3 \cdot 4\text{SiO}_2 \cdot \text{H}_2\text{O}$	{ 0.048 Low
Montmorillonite.....	$\text{Al}_2\text{O}_3 \cdot 5\text{SiO}_2 \cdot (\text{Mg}, \text{Ca})\text{O} \cdot ?\text{H}_2\text{O}$	0.025
Steargillite (gouge).....	$\text{Al}_2\text{O}_3 \cdot 4\text{SiO}_2 \cdot \text{H}_2\text{O} + ?\text{H}_2\text{O}$	Low?
Beidellite (gouge).....	$\text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot ?\text{H}_2\text{O}$	0.041
Leverrierite ¹	$\text{Al}_2\text{O}_3 \cdot 4\text{SiO}_2 \cdot 6\text{H}_2\text{O}$	0.028
Sericite.....	$3\text{Al}_2\text{O}_3 \cdot 6\text{SiO}_2 \cdot \text{K}_2\text{O} \cdot 2\text{H}_2\text{O}$	0.033

¹ Leverrierite has small variable contents of K_2O (Corbett, Am. Jour. Sci., 5th ser., vol. 10, p. 251, 1925).

VARIETIES OF CLAYS AND SHALES

(Mostly lutites)

A. Textural and structural varieties.

Loose or Friable	Well Cemented
1. Fine clay and shale.....	(and argillite and slate)
2. Coarse clay and shale.....	(and argillite and slate)
3. Silty clay and shale.....	(and argillite and slate)
4. Sandy clay and shale.	(and argillite and slate)
5. Gravelly clay.....	(and argillite and slate)
6. Boulder clay or till.....	Tillite
7. Concretionary clay and shale.....	(and argillite and slate)
8. Fossiliferous clay and shale.	

The following are also used as structural prefixes: massive, laminated, varved, slickensided, folded, faulted, foliated, mud cracked, rain printed, "bubble prints," rippled.

B. Mineralogic varieties.

Red, yellow and brown = ferruginous clay or shale.

Black and green = carbonaceous clay or shale.

Blue mud, commonly pyritic.

Calcareous shale.

Carbonaceous shale	{	oil shale.
		oil rock, bituminous shale.
		paraffine shales.
		bone coal.

Siliceous (usually sandy) clay or shale.

Diatomaceous shale.

Gypsiferous clay or shale.

Micaceous shale.

Ashy and tuffaceous clay or shale.

Bentonite (waxy luster; swells enormously).

Gouge.

Fire clays.

Kaolin, white with much kaolinite.

Bauxitic clay.

"Copper-bearing" shale.

C. Special terms.

1. (A series of special terms based on uses.)

Brick clay, pottery clay, modeling clay, and many others.

2. (A series of terms based on origin.)

Glacial clay, fluvial clay, lacustrine clay, marine clay, residual clay, and others. The criteria of origin are not always clear, however, when the deposits are being studied and named.

a. **Saprolites**, residual clays transported no more than a short way, forming a surface mantle of soil. Average saprolite is impure, usually red.¹ "Terra rossa" and "kunkar" are formed from limestone or contain limy concretions. Some men use the term laterite as originally defined for these same residual soils.

b. Concretionary residues from which silica has been leached even from aluminous compounds, form *bauxite* and *iron oxides* more or less impure but with little silicate; the masses form hard surface crusts in tropical regions and are called laterite by some geologists,² (see pages 294 and 297).

c. **Loess** is wind blown silt or silty clay, with commonly little stratification.

3. **Fire clay** is a term used by coal miners for clay beneath the coal regardless of its character.

4. **Hardpan** is boulder clay that is difficult to excavate.

5. **Gumbo** and *gumbo-till* are particularly sticky when wet.

6. (Age terms and terms showing the locality of production tell very little as to petrographic character.)

Notes on the Clays and Shales.—Argillite and slate may properly be classed as metamorphic rocks, but they are included here because they are commonly considered as part of the sedimentary sequence in any field mapping and in geologic sections.

Diagenesis (page 262) may develop in clays a certain degree of induration, some concretions and some mineralogic readjustments. It is likely that colloidal material in the sea may coagulate and soon become crystalline. Glauconite may form at this stage. Sericite is said to develop early.

Concretionary clay is usually residual clay in which rearrangement has developed the structure in the clay substance. Quite different are the clays that bear concretions of other minerals, usually carbonates of lime or iron (limonite after weathering) (see page 294).

Lamination. The normal sequence of grain in a bed deposited as a result of one climatic impulse is a gradation from coarse grain at the base to fine grain at the top. This is a result of the slower settling of fine material as well as a slower supply in the part of the cycle just before the next impulse. The bed, if thin, may be called a lamina, especially if the gradation is too small to be visible. The large laminae showing gradation of grain are called "varves" (Figs. 164 and 167).

There is also, however, a tendency for clays to be laminated or have a cleavage along the bedding, even if the material seems to be uniform.

Mechanical analyses show that most clays of commercial application have more grains of silt size than of clay size.

Clays have commonly a high porosity, but the grains are so small that the openings between grains are subcapillary, and hence clay formations are not favorable to the circulation of water or oil in the ground. In

¹ HARRASOWITZ, H., *Laterite*, Fortschritte Geol. u. Palaeont., Berlin, 1926.

² See SCRIVENOR, J. B., *Geol. Mag.*, vol. 67, pp. 24-28, 1930.

shales and still more in argillite and slate the porosity is reduced by recrystallization and pressure (see page 337).

The term fire clay for clays beneath coal is based on the fact that some such clays are refractory. It has been found, however, that many clays with coal are not refractory.

Probably most black shales are carbonaceous, but some are blackened by iron sulphide (see further readings).

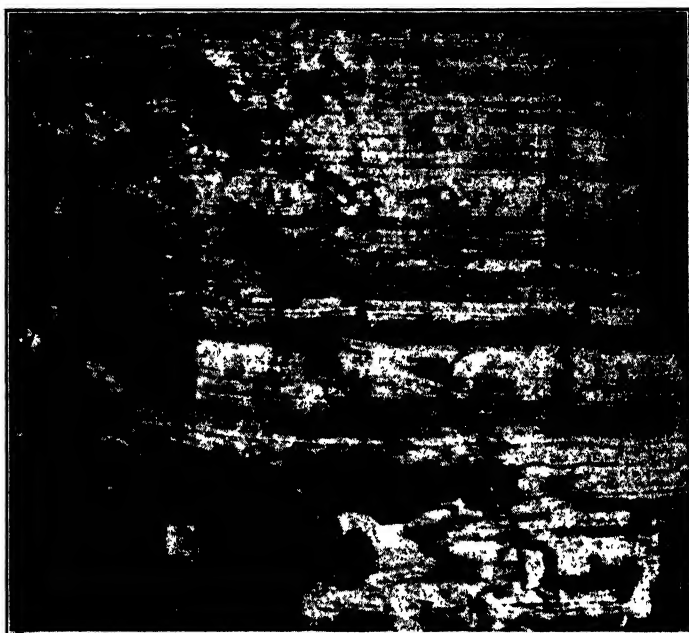


FIG. 167.—Laminated clay, Wrenshall, Minnesota. An example of varved clays formed in glacial lakes.

The water-laid fluvial and lacustrine clays are distinguished with difficulty from the more-abundant marine clays (see the criteria of places of deposition, page 322). Deep-sea ooze, consisting largely of altered dust and ash, makes the "red clay" of a few exposures, but deep deposits are rarely elevated into continental masses.

Residual clays having formed from various rocks may have residues of any of the original minerals as impurities. The pure well-leached residues are kaolin. Especially thorough leaching may leave only the hydrous oxides of iron and aluminum. Some leached residues develop oölitic or concretionary textures. Ferruginous residues of this texture are sometimes called lateritic iron ores, but the process may not necessarily have been thorough enough to leach silica from the aluminous compounds. A slightly transported aggregate of concretions may form a "pseudolaterite."

Minerals in Clays.—The hydrous aluminum silicates are numerous (page 269) and the customary reference to most of the clays as containing kaolinite, is simply a convenience based on lack of knowledge, for there is much difficulty in an accurate determination (*x*-ray methods may be of some assistance). Several occurrences of crystalline kaolinite justify the recognition of that substance as a definite mineral and possibly dominant in some clays. The several silicates differ in their physical properties, and possibly most weathered residues are colloidal rather than crystalline.

Differing from these products of decomposition or "rock rot," some glacial clays consist of pulverized rock matter with very little decomposition, called "rock flour."



FIG. 168.—Two equal blocks of bentonite, the one at the left wet with water, the one at the right remaining dry.

The minerals beidellite and montmorillonite are characteristic of bentonite, which forms from volcanic ash beds.¹ Bentonite swells enormously when wet (Fig. 168) and forms a milky suspension in water.

TABLE XIII.—AN ESTIMATE BASED ON CALCULATION FROM ANALYSES (PAGE 264), OF THE MINERALS IN THE FINER FRACTIONS OF TWELVE CLAYS OF VARIOUS ORIGINS

Estimated minerals	Silt, per cent	Coarse clay, per cent	Fine clay, per cent
Kaolinite and clay minerals	7.5	17.0	23.2
Sericite and paragonite	16.6	21.2	22.1
Quartz	36.7	19.3	13.1
Chlorite and serpentine	8.2	10.3	7.3
Limonite, hematite (and pyrite)	3.0	5.5	8.0
Calcite and dolomite	10.5	7.5	5.7
Feldspars	12.6	7.2	7.3
Zeolites	3.0	7.5	6.9+
Titanite and rutile	1.7	2.0	1.7
Carbonaceous matter	0.2	0.9	0.6
Moisture	0.9	1.3	4.1

In particular clays, the minerals identified are commonly only those in the coarser fractions. Almost any mineral or rock may be present

¹ See DAVIS and VACHER, U. S. Bur. Mines Tech. Paper 438, 1928.

(see the list of minerals in sands, pages 474 to 478). Soluble minerals are least likely. In shales and slates a few metacrysts and concretions are easily recognized as pyrite, graphite, chlorite, calcite, gypsum or others. In the main bulk of fresh material, however, the estimates of minerals in the finer portions are based on difficult work with a microscope, and rough estimates from analyses, and so much of the clay may be colloidal that the work can not be accurate. Sericite, hydromica, rutile, kaolinite, etc., are reported. The nature of the magnesian and other minerals is quite uncertain.

Between pure clay and limestone there are all possible gradations. Calcareous clays with 25 per cent limy material are so hard as to appear like limestone. The division between limestone and shale is therefore probably made most commonly at about 20 per cent of limy material. Sandy material, however, up to 75 per cent is so inconspicuous in a mixture with clay that such mixtures are usually classed as shale.

SANDS AND OTHER MECHANICAL SEDIMENTS (NOT CLAY AND LIMESTONE)

(Largely hydroclastic)

A. Textural and structural varieties.

	Loose or friable	Well Cemented
<i>Rudites</i>		
Gravel	Conglomerate	Conglomerate quartzite
Scree and talus	Breccia	
Glacial till		Tillite
<i>Arenites</i>		
Coarse sand	Grit	Quartzite grit
Sand. Sandrock ¹	Sandstone	Quartzite
Silt. Siltrock	Siltstone	
Sandy clay	Shaly sandstone	Slaty quartzite
Pebbly sand	Pebbly sandstone	Pebbly quartzite
<i>Lutites</i>		
Siliceous ooze		
Tripoli		Novaculite

Other structural features may be included as prefixes to the name, such as massive, stratified, bedded, laminated, reeded, cross bedded, concretionary, fossiliferous, ripple marked, mud cracked, or preserving rain prints, foot prints, swash marks, beach cusps, ice marks, bubble prints, clay galls, mud lumps, fulgurites (see pages 323 to 325).

Intraformational conglomerates have pebbles incorporated in new strata nearly contemporaneous with original beds. Many have flat pebbles (Figs. 169 and 204D) lying in the plane of the bedding.

¹ Probably loosely cemented sands have more often been called "friable sandstone" than "sand rock," the term proposed by Tieje, Jour. Geology, vol. 29, p. 655.

Edgewise conglomerates have flattened pebbles transverse to the main bedding.

Breccias may be distinguished as crackle, mosaic and rubble, according to the extent of disturbance in the position of the fragments, which is greatest in rubble breccias.

B. Mineralogic varieties.

"Pure," highly quartzose or siliceous sandstone; white sandstone.

Arkose = feldspathic sandstone.

Graywacke; contains enough black and gray grains to give the rock a gray color.

Glaucconitic sandstone = "Greensand."

Magnetitic sandstone = "Black sand."

Tuffaceous sandstone (see tuffs in section on Igneous rocks; some water laid).

Brown, yellow and red = ferruginous sandstone, etc.

"Brownstone."

Argillaceous or shaly sandstone.

Calcareous sandstone.

Loam = argillaceous and silty sand.

Marl; mixed minerals.

More rarely the dominant sand grains may be hornblende, olivine, dolomite, etc.

Breccias and conglomerates are named also from rock fragments, as granite conglomerate, limestone conglomerate, diabase breccia, quartzite breccia.

It is desirable to establish a distinction between the minerals of the grains and those of the cement. Possibly an adjective may indicate the cement and a mineral prefix the nature of the grains, but it can not be said that the usage is generally understood and accepted. It may be illustrated by:

Ferruginous quartz conglomerate.

Calcareous glauconite-quartz sandstone.

Calcareous trap breccia.

Dolomitic greensand marl.

Ferruginous arkose (sandstone).

C. Special and locality terms.

(Age terms are much used but are not very good signs of petrographic character.)

(Terms indicating origin, similarly.)

Flagstones, thin-bedded sandstones.

Itacolumite, flexible sandstone.

Buhrstone, cherty, porous, etc.

"Red beds," shaly red sandstone.

Hardpan, a conglomerate or gravel cemented with limonite, etc., found in placer digging.

Case-hardened sands have limonite or other minerals as a surficial cement or desert varnish.

Certain quarries with large commercial production are widely known for sandstones of definite petrographic character, as, for example, the Berea grit.

Notes on Sands and Other Mechanical Sediments.—Conglomerates have been given detailed study¹ as to field relations, structure, pebbles, matrix and color. These features may serve to distinguish conglomerates of different origins: (1) marine; (2) lacustrine; (3) estuarine; (4) fluvial (fluvialite); (5) crush; (6) glacial. Glacial conglomerates are said to



FIG. 169.—A flat-pebble conglomerate in the midst of the iron-bearing formation of the Mesabi Range, Minnesota. The pebbles are much like the matrix. About one-half natural size.

differ from others in absence of weathering of the matrix—Na and Ca are not reduced as they would be by weathering. The recognition of lithologic types in the pebbles of a conglomerate or breccia often enables the geologist to trace the formation to its source and make out the geography of the region in past times.

“Pseudo conglomerates” are not truly mechanical sediments but show concretions or “augen” or balls from spheroidal weathering that resemble pebbles.

Breccias may be distinguished by origin as talus, residual, friction and igneous breccias. Subaerial, subaqueous and endolithic breccias may have similar significance.² The criteria for their identification involve both fragments and matrix but are not always conclusive. Scree is the material in rock slides on steep slopes.

Bedding in the coarser rocks may be obscure (Figs. 170 and 173). Even in some sandstones there may be beds many feet thick that seem

¹ MANSFIELD, G. R., Origin and structure of the Roxbury conglomerate, Harvard Col. Mus. Comp. Zool., vol. 49, Geol. ser., vol. 8, pp. 91-271, 1906. A. C. LAWSON, California Univ. Pub., vol. 7, p. 325, 1913; Geol. Soc. America, vol. 23, 1912.

² NORTON, W. H., A classification of breccias, Jour. Geology, vol. 25, pp. 160-194, 1917. See also LEITH, “Structural Geology,” rev. ed., pp. 106-110, 1923.



FIG. 170.—Block of conglomerate, about one-half natural size.



FIG. 171.—A friction breccia with brown calcareous cement. About one-half natural size.

devoid of internal structure. From these thick beds there are all gradations to beds a fraction of an inch thick. Laminae differ from each other in little but color, whereas beds differ in texture and composition. Detailed studies of the cross bedding (Figs. 174 and 175) and ripple marks



FIG. 172.—Sketch of a bed of conglomerate with imbricate arrangement of pebbles Little Vermilion Lake, Ontario. (After Pettijohn, *Jour. Geology*, vol. 38, p. 572, 1930.) Shaded pebbles are metamorphic rocks; unshaded are granite. If this arrangement is a result of stream action the pebbles dip upstream.

(Fig. 176), etc., may give a clue as to the conditions of deposition (see pages 323 to 325). Once the sandstone is cemented it is likely to have joints crossing the bedding (Fig. 177).

Textural and structural variety names include one or two names to indicate the shape of fragments, as for example, breccia for angular, and

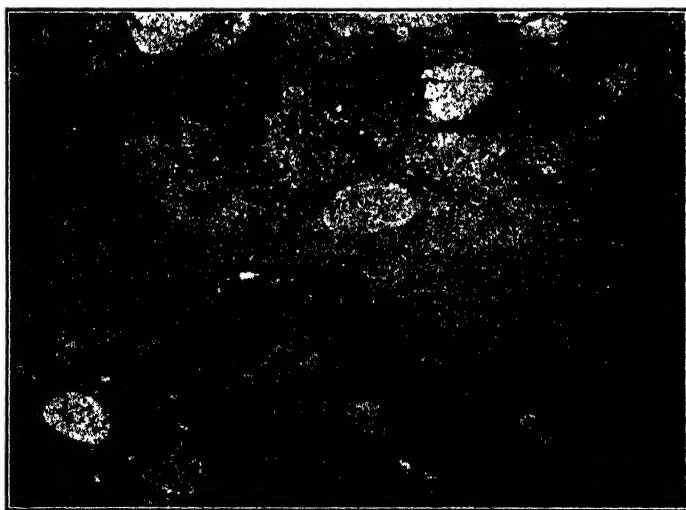


FIG. 173.—A conglomerate near Sioux Lookout, Ontario. (Photograph by F. J. Pettijohn.) The bedding is obscure except where sand is locally interbedded.

conglomerate for rounded pebbly rocks; but for sands this refinement of nomenclature is not yet standard usage. "Grit" is suggested for sharp sand as distinct from rounded sand, but in America the term is more used for coarse than for sharp grains. The names of sandstones may be prefixed by some term or phrase to indicate the shapes of grains (Figs. 178 and 179).



FIG. 174.—Cross bedded sandstone under a basalt lava flow. Grand Portage, Minnesota.

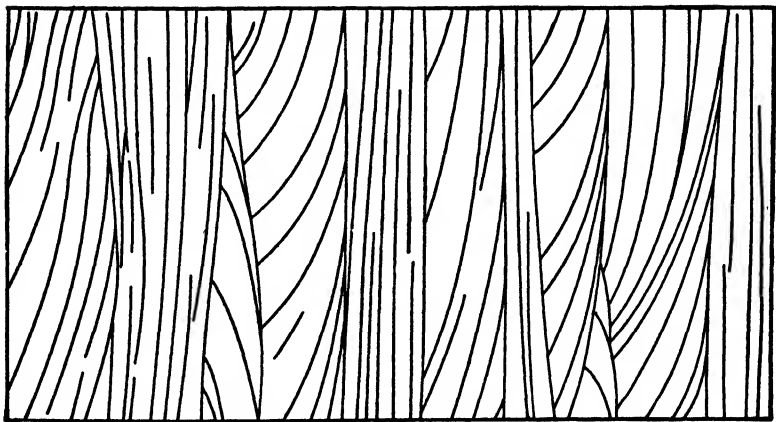


FIG. 175.—Diagram of cross bedding in a sandstone, showing the features by which the tops of the beds can be determined after tilting. Tops of the beds are here at the left.

Forms of fragments have been studied in detail by a ratio involving the radius of curvature of the sharpest edge and the average radius (page 315). Pebbles are commonly flat ("discoid") on beaches or in shallow water; rounded ("spheroidal") in streams; subangular, and striated on



FIG. 176.—Sketch of ripple marked sandstone.



FIG. 177.—Coordinate joints and bedding in a graywacke sandstone, Cook County Minnesota.

flattened cut faces (faceted) in glacial till, etc. "Dreikanter" are pebbles cut by wind-blown sand so as to show three facets and three limiting ridges; einkanter with one crude ridge are more common. Residual pebbles, like cherts from limestone, are apt to be very irregular. The original form and structure of the fragments have surprisingly persistent control. "Quick" sands are not extra well rounded, only extra wet. Grains are divisible into about four different grades; (1) angular, (2) subangular, (3) subrounded, and (4) rounded (Fig. 178).

Wentworth's mathematical ratio of the radius of curvature of the sharpest corner to the mean radius has led him to the conclusion that few grains ever become well rounded; but the four divisions here given are readily estimated by most men with fair agreement. A roundness ratio of 0.2 might be acquired in 40 or 50 miles of transportation.¹

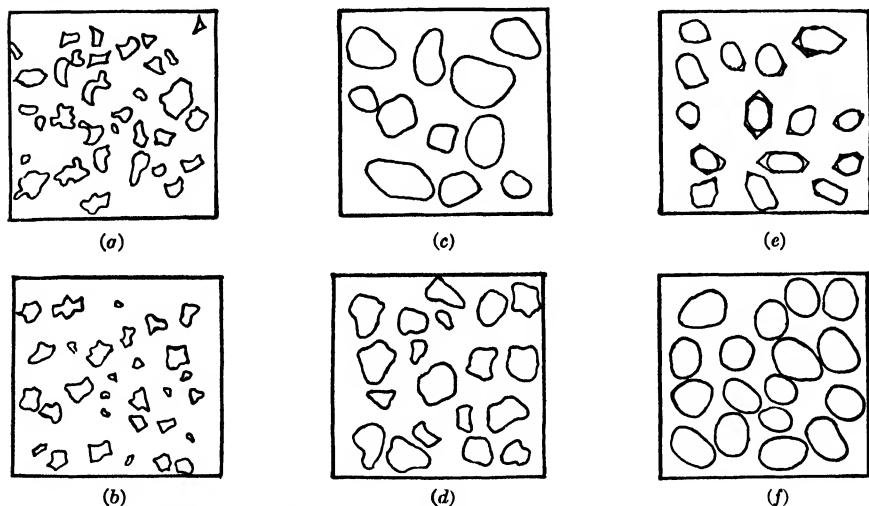


FIG. 178.—The forms of sand grains as seen under the microscope, (a) Sharp angular, residual. (b) Sharp angular, glacial. (c) Sub-rounded, dunes on shore. (d) Sub-angular, beach. (e) Rounded but with secondary crystal growth. (f) Well rounded, wind blown.

The fragments in sediments are not always easily distinguished from spots in igneous rocks, and criteria are here tabulated for the distinction of fragments from phenocrysts and amygdulcs.

CRITERIA TO DISTINGUISH FRAGMENTS, PHENOCRYSTS, AND AMYGDULCS

(See also metacrysts and phenocrysts, page 356)

Forms.

Phenocrysts have crystal forms, possibly corroded, parallel opposite sides, regular arrangement of faces and angles.

Fragments, irregular to rounded.

Amygdulcs, spheroidal to ellipsoidal to cylindrical, almond shaped.

¹ WENTWORTH, C. K., A method of measuring and plotting the shapes of pebbles, U. S. Geol. Survey Bull. 730 C, pp. 91-111.

Minerals.

Phenocrysts commonly quartz, feldspar, hornblende, augite, olivine or biotite; some may be replaced forming pseudomorphs.

Amygdules commonly quartz, calcite, zeolites, chlorite, epidote.

Fragments commonly vein quartz, granite, quartzite, trap rock, limestone, other igneous and metamorphic rocks. Fragments may be of glass, or of cold-water minerals, or of more than one kind of mineral; phenocrysts and amygdules rarely show such features.

Matrix.

Phenocrysts and amygdules lie in a groundmass of felsitic igneous rock. Fragments usually lie in a matrix of sandy or shaly sediments, or a cement of minerals deposited by cold water. The groundmass may show flow structure around phenocrysts and amygdules, but the matrix of fragments shows bedding rather than flow.

Most sands by far are composed dominantly of quartz; when the term sand is used without qualification, a sand consisting largely of quartz is

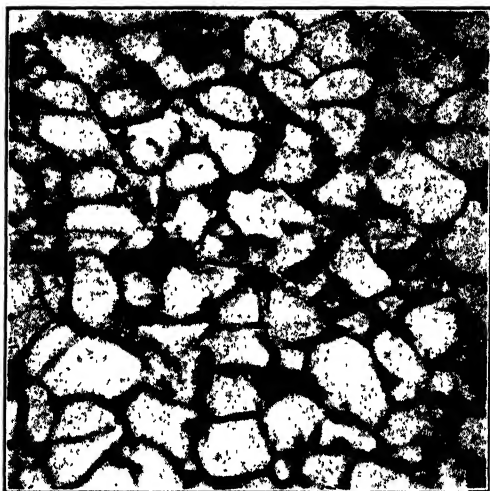


FIG. 179.—Photomicrograph of sandstone, angular grains in a hematite cement. $\times 50$.

meant. Similarly, arenaceous clay is assumed to have quartz sand grains. The term "arenite," however, is more generally understood to be wholly textural in meaning.

Such terms as marine, terrestrial, lacustrine, etc., are not of much descriptive value, because based too largely on uncertain interpretations (see the list of further readings). *Atmoclastic*, *hydroclastic*, etc., may be a little better.

Quartzites and equally metamorphosed sediments, though commonly called metamorphic, are included here, because they are listed in most stratigraphic sections as a part of the sedimentary series.

Silt, here defined as fine sand, is a popular term for the muddy mixed sediments carried by a river. Alluvium may also be used as a rock name for deposits of river débris.

Loam is variously defined, but it is a serviceable term in petrography for the mixture of some silt and clay with dominant sand.

Diagenesis may develop in sands and mechanical sediments such features as concretions and cement before the next beds are deposited. Glauconite probably develops at this stage.

Later cementation may be diffused or concretionary or in a few cases crystalline. The cements of sandstones may form 25 per cent of the total mineral content. In the gradation of sand to clay, about 25 per cent of clay makes the rock very shaly in appearance, and the more shaly mixtures

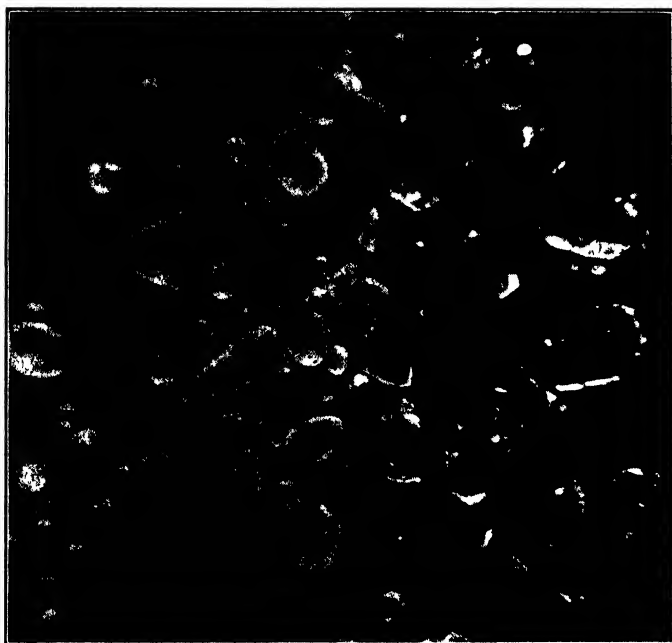


FIG. 180.—Photomicrograph of sand grains on a black background. Half the field shows well polished grains, and the other half, frosted grains. $\times 60$.

are likely to be called shales. On the other hand, limy sandstones may have more than 50 per cent calcite and still look very sandy. The division is probably made at about 20 per cent quartz.

The common cements of sand are quartz, limonite, calcite, and clay; less often opal, chalcedony, hematite, other carbonates. Many ferruginous sands have bleached spots resulting from local deoxidation. In flexible sands the cement is said to be muscovite. Rarer cements include barite, pyrite and other sulphides, zeolites, chlorite, gypsum, epidote, talc, serpentine, hornblende, mica, phosphates and gibbsite. Of commercial interest, several ore minerals may be found as cements of sands; malachite and other copper minerals are perhaps most abundant, but galena, carnotite, and others are known,

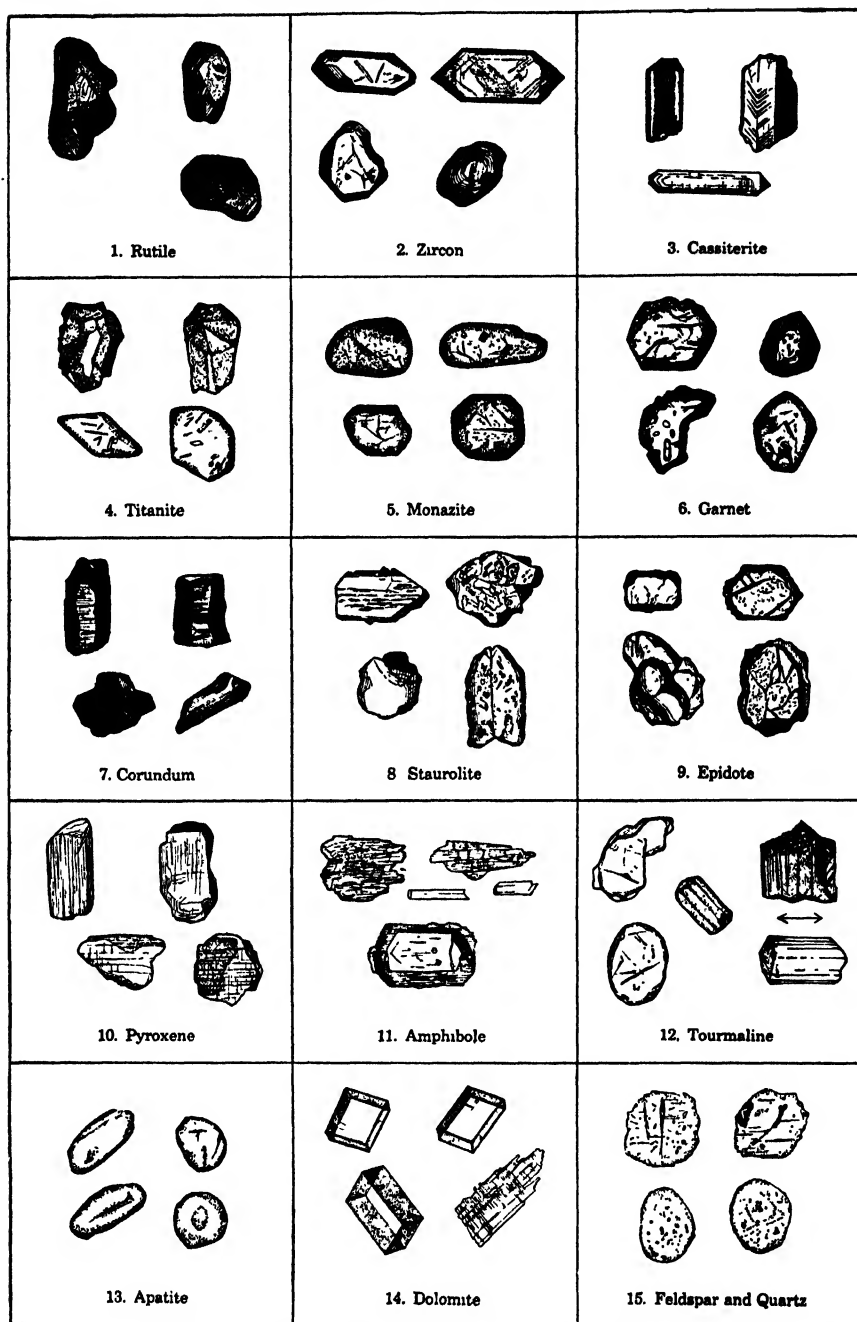


FIG. 181.—Sketches of the microscopic appearance of minerals separated from sands and mounted in balsam; numbered in the order of decreasing index and relief. $\times 20$. 1. Rutile, Rusthall, Kent. 2. Zircon, Alabama. 3. Cassiterite, Cornwall. 4. Titanite,

Cement deposited in crystallographic continuity with the sand grains is spoken of as secondary growth, and is common in quartzites. The rocks with finer quartz, chalcedony or opal cements may be termed siliceous rather than quartzitic sandstones. Cement in large crystals enclosing many sand grains is illustrated by "sand calcites" and "sand barites." The cement that is concentrated on the outside of sandstone pebbles as a desert varnish probably results from evaporation¹ though organic secretions by lichens may be involved. Small tubes of fused silica in sands are fulgurites, and are attributed to lightning.

Minerals of Sand.—Attention has been called to the dominance of quartz in sands, but some 200 sand minerals are on record. Locally some one of these may form a high per cent of the rock. Most petrographic studies of sand cover chiefly the detailed features of quartz, or the proportion and nature of the other minerals.

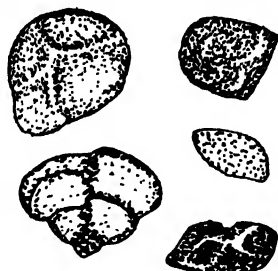


FIG. 182.—Glauconite separated from a sand. $\times 50$.

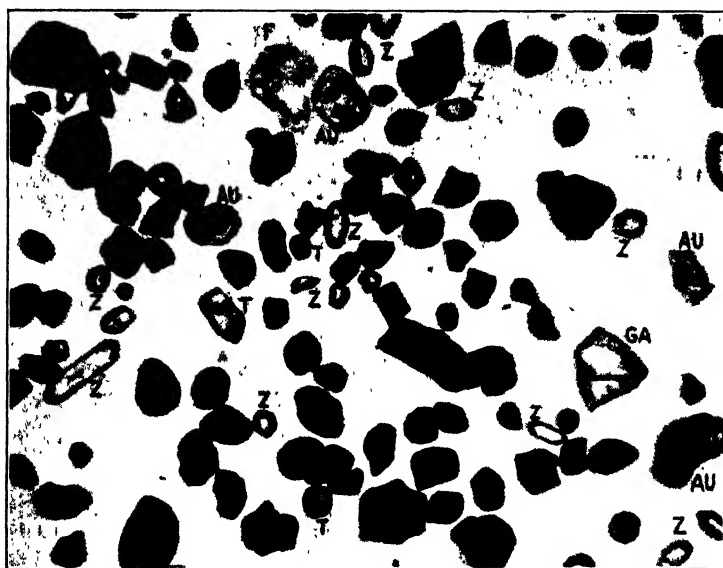


FIG. 183.—Mixed sand from a gold placer concentrate, Alabama. Au, gold; Z, zircon; T, tourmaline; Ga, garnet; black grains are magnetite. One large feldspar (F) (a mineral of low specific gravity) contaminates the concentrate. $\times 100$.

¹ LAUDERMILK, J. D., Origin of desert varnish, *Am. Jour. Sci.*, 5th ser., vol. 21, pp. 51–66, 1931.

New England. 5. Monazite, Brazil. 6. Garnet, Connecticut. 7. Corundum, Carolina. 8. Staurolite, North Africa. 9. Epidote, Anglesey. 10. Pyroxenes, Minnesota. 11. Amphiboles, Minnesota (note pleochroism). 12. Tourmaline, Cornwall (note pleochroism). 13. Dolomite, Wisconsin. 14. Apatite (Yorkshire). 15. Quartz (clear) and Feldspar (dusty), Minnesota.

Quartz grains differ internally as to inclusions and as to extinction, which is commonly wavy. Wavy extinction results either from strain from metamorphism in the earlier history of the quartz or from flamboyant growth in a vein—a more common occurrence than is generally realized (Figs. 194 and 226). A later strain effect near faults may be distinguished by the fact that nearly all grains are affected, whereas the earlier effects are localized in a few grains, mixed with other grains showing no strain.

The external features are size, shape, polish or frosting (Fig. 180), coatings and stains.

The minerals other than quartz are studied mostly by a separation of the heavy residuals (see pages 264 and 265 and Table VII). This leaves the light minerals, feldspar, glauconite (Fig. 182), muscovite, cordierite, serpentine, etc., commonly disregarded unless there is a thin section.

Probably the next minerals in abundance after quartz are the feldspars, characteristic of arkose and graywacke sands. Arkoses are commonly coarse (grit) and have bleached mica as an accessory (see page 308).

The commonest heavy residuals, zircon, garnet, tourmaline, magnetite and apatite, are so largely derived from grains that were originally small that they are studied mostly in the grains separated between 0.5 and 0.05 mm. in size. The smaller sizes may have the same minerals but less easily studied. These commonest of heavy residuals are so nearly universal that their presence is seldom of value in petrographic studies or correlation. The peculiarities of the grains and their relative abundance are more characteristic in some formations. The student should be thoroughly familiar with typical material of these minerals, so that any peculiarities and any other minerals are readily noted (see Figs. 181 and 183).

Other heavy residual minerals, if abundant in a concentrate, are more likely to be local and to be of considerable value in identifying formations and in studies of the source of the sediments.

LIMESTONES

By strict definition limestone is a bedded deposit consisting chiefly of calcium carbonate and yielding lime when burned. In a broader way limestones include dolomites, in which calcium carbonate may be less than half, and it is possible that some pre-Cambrian highly carbonated slates or greenstones have been classified as limestones.

The limestones form a complex group, differing not only in the way the carbonate is precipitated but even more in the processes of reworking and diagenesis. The textural series based on deposition of organic remains more or less crushed is not so universal as was once supposed;

but many limestones have organic remains reworked into calcareous sand or mud and scattered in a matrix that may be precipitated by microorganisms or by chemical action.

The textures are in marked contrast with those of mechanical sediments for several reasons, chiefly that relatively few limestones are built from fragments of preexisting rocks. Some are precipitated from solution in crystalline form or recrystallized from fragmental material, but a large part of most limestones consist of fragments of fossils, ranging in size



FIG. 184.—Coquina, Florida. A shell limestone. Natural size.

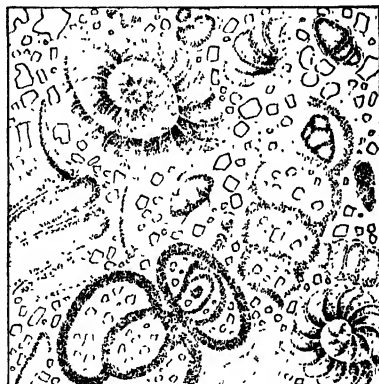


FIG. 185.—Sketch of a thin section of Pleistocene limestone, California. $\times 25$.

from microscopic algae to coral reefs. The resulting forms and structure are therefore quite distinctive (see Figs. 184 and 185).

VARIETIES OF LIMESTONES

A. Textural and structural terms and degree of cementation.

1. Normal texture series, biogenic.

Loose or friable

Well cemented

Calcirudites.

Shells and other remains

Shell limestone { Coquina
Coral limestone
Algal limestone,
etc.

Brecciated limestone

Fossiliferous limestone

Limestone conglomerate

Limestone breccia

Calcarenites.

Calcareous sands Granular lime rock Granular limestone

Calclutites.

Crystalline limestone

Calcareous ooze

Chalk

Fine grained limestone

Lithographic stone

Marl

Marbles

2. *Limestones of mixed origin.*3. *Precipitated, hydrogenic.*

Öolitic and pisolitic limestone.

Concretionary limestone (including cone-in-cone).

Limestone with concretions of other minerals.

Stalactitic limestone.

Mexican onyx.

Travertine (botryoidal, spherulitic, platy, concentric, etc.).

Caliche (in part).

Vein calcite.

4. *Other structural terms.*

Structural terms include also massive, stratified, and even (when shaly) laminated, cross bedded, stylolitic and porous especially if dolomitic. Fossils have various internal structures.

B. *Mineralogic varieties.*

"High-calcium" limestones.

Magnesian limestones or dolomitic limestones.

Dolomite limestones.¹

Argillaceous limestone (cement rock, hydraulic limestone).

Arenaceous limestone (usually siliceous).²

Ferruginous limestone (red, brown, etc.).

Carbonaceous limestone (bituminous).

Glaucconitic limestone.

Cherty limestone.

Pyritic limestone.

Phosphatic limestone.

Tuffaceous limestone (variety of mineral and rock fragments).

C. *Special terms.*

Age terms are common but have little petrographic significance.

Certain quarries produce stone of standard character, widely known; as, for example, Bedford oolite.

Terms based on origin are seldom safely applied. For example, "fresh-water limestone" is distinguished from the more abundant marine limestone by fossils rather than by the more strictly petrographic features.

Notes on Limestones.—The table shows "shell limestone" or "coquina limestone" for rocks in which most of the material is in recognizable shells; "fossiliferous limestone" for those in which fossils are less common and in a fine matrix. It may be well to name the fossil if one type largely dominates (see below). The deposition of fragments of calcareous rocks as mechanical sediment hardly needs discussion. Wave and wind action produce and transport and deposit calcareous sands and muds. The resulting textural types of limestone as shown in thin sections have been described in detail and illustrated,³ entirely aside from the commonly associated organic forms. The noteworthy points include the size of

¹ Dolomite is a mineral name and should not be used unmodified for a rock.

² All the minerals of sands and clays may be intermingled.

³ See HOWELL, J. V., Pre-Permian Paleozoics of the Wichita Mountain area, Am. Assoc. Petroleum Geologists Bull., vol. 6, pp. 415-420, 1922.

grain and its uniformity, shapes, twinning, deformation and the siliceous or argillaceous cement and its abundance.

The accessory and varietal minerals of limestone differ greatly in their textural relations in the rock. Dolomite is normally a disseminated replacement, clay is mechanically associated and chert is a concretionary replacement.

Marl is a term of varied usage, nearly always implying an earthy mixture of minerals. In some places the main materials may be quartz and glauconite sands, in others clay and calcite. From the calcareous clays the name marl has come to be used for earthy calcareous deposits in lakes and swamps even though nearly free from clay. Marl is good cement material in the north-central part of the United States.

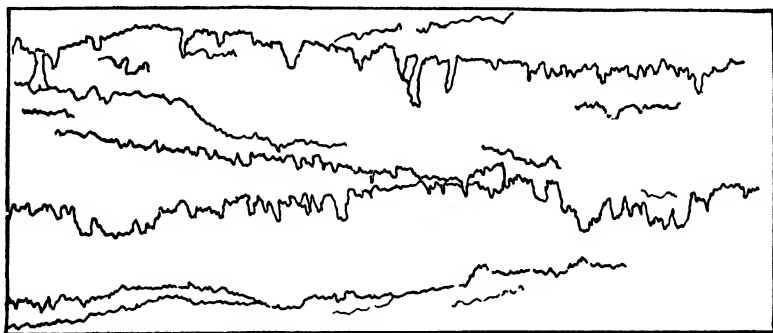


Fig. 186.—Stylolites in Tennessee "marble," a crystalline limestone. About one-half natural size.

Limestones, more than other common sediments, are especially subject to diagenetic changes, probably because of the solubility of the material. Calcite may be replaced by phosphate, or dolomite, or hematite before being covered under later sediments. Probably a large proportion of limestones become somewhat crystalline and cemented before deep burial. Calcite in most limestones seems to recrystallize too readily to retain much intergranular space after burial under other sediments. Some chalks are more permanently porous. The stylolitic and cone-in-cone structures are also probably related to the solubility of the rock (Fig. 186).

Oölitic, pisolitic, and concretionary limestones are spheroidal or ellipsoidal structures built (Fig. 187) of concentric layers around a nucleus which may be a fossil, a sand grain or even some inconspicuous grain of the same mineral as the rock. The layers ultimately comprise a volume far greater than the volume of the nuclei. The distinction of the three terms is based largely on the size of the finished product, oölitic about the size of pin heads, pisolites about the size of a pea, and concretions larger. Large calcareous concretions in other rocks than limestone may lose most of the concentric structure characteristic of the oölitic.

The precipitated hydrogenic limestones may be partly formed by microscopic organisms, but the forms are not conspicuous. Travertine is porous and deposited partly on vegetation (Fig. 202). Onyx is more compact and banded.

Fossils are too numerous to cover thoroughly in a course in petrography. Several forms are so large that a thin section or even a hand specimen of reasonable size is too small for a clear illustration. The forms best studied in section are probably foraminifera, but algal and bacterial structures also can be seen with the microscope in some rocks (see Fig. 185).

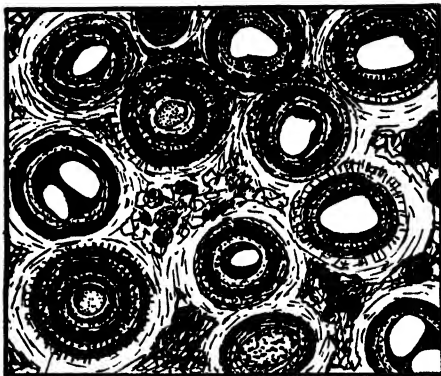


FIG. 187.—Sketch of a thin section of oolite, Minnesota. Each spheroid is likely to have a nucleus or core of sand, or a fossil fragment. The thin section, however, may not cut the nucleus. $\times 25$.

As characteristics of the organisms of common limestones may be mentioned:

Algae produce some oolites (Fig. 187) and concentric structures of larger size, cryptozoons. Probably much of the algal limestone is structureless. Bacteria may form other minute structures.

Foraminifera occur entire, most of them being 1 mm. or less in diameter. Some chambers may be detached; some tests may be filled with radial calcite, others with glauconite.

Corals commonly have a structure consisting of fibres of calcite or aragonite.

Crinoid stems may be dominant in some limestones, each cylindrical joint behaving as a single crystal of calcite, but with the form and canals of the organism marked by dusty areas in the crystal.

The *crustacea* have shells of chitin, with a little calcite in fibres perpendicular to the surface, and in the direction of the optic axis.

Brachiopods have calcite fibres and laminae, part parallel and part at various angles to the surface of the shell, and with inclined extinction.

Pelecypods are somewhat similar, but variable, and the shells are inclined to be stouter.

Gastropods and *cephalopods* have shells of aragonite more or less replaced by a mosaic of calcite.

The Minerals of Limestones.—Dolomites are distinguished from calcium limestones by their slight effervescence in cold acid. Many dolomites are porous on account of their origin as a replacement of calcite (see page 338). One further feature is locally useful as a distinction, namely, the dolomites are commonly a little more ferruginous than limestones and the fresh gray rock looking like limestone, weathers to a

buff or pinkish brown that is rare in the non-magnesian limestones. The several carbonates are not easily distinguished in thin sections, but an estimate is commonly made on the basis of the multiple twinning of calcite (Fig. 234) and the more nearly euhedral rhombs of dolomite, (Fig. 188). Stains may be applied if the textural relations of several carbonates are to be studied, and the indices of refraction in oils serve to distinguish not only these common carbonates but others in the group. Few limestones show crystals intermediate between dolomite and calcite, but dolomitic limestone is a mixture of true dolomite and calcite. Aragonite may be seen in shells and as needles in some precipitated carbonate; it is not magnesian.

The division between arenaceous limestone and calcareous sandstone might be logically set at 50 per cent of each, but so few limestones occur with 10 to 20 per cent sand and so many more with 40 to 60 per cent sand that the limit may be placed at 20 per cent sand. It is logical, also, because quartz sand grains become very noticeable, if they make up 20 to 40 per cent of a mixture.

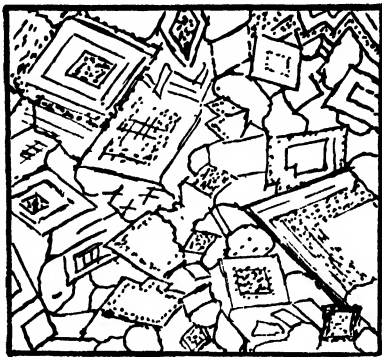


FIG. 188.—Sketch of a thin section of dolomite showing the characteristic rhombic crystals and zoning. $\times 100$.

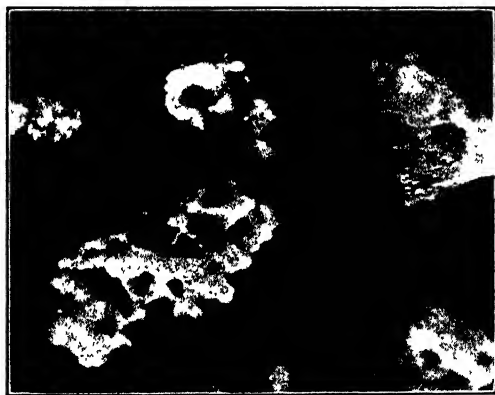


FIG. 189.—Photomicrograph of the cherty residue left from the solution of a dolomite. Dolomite rhombohedra dissolved out leaving "dolocasts." $\times 50$.

The mixtures of limy matter and clay, on the contrary, retain the characters of limestone until the percentage of clay is very great. Kaolin may be disseminated in many limestones as dust, almost indistinguishable except by analysis from fine grains of carbonate.

Some fossils and pebbles in limestones are dark from disseminated organic matter, pyrite and manganese oxides. Many pebbles, including some of these darkened grains, are phosphatic, with collophanite, dahllite or some other phosphate minerals.

The heavy residual minerals from limestone are rarely separated by simple panning or by the use of a heavy liquid. Much more commonly the first treatment is the removal of the carbonates. The "mass characters" of the residue from solution of most limestones are sufficiently characteristic to serve all purposes of description and correlation, but if more detailed work is needed the heavy liquid separation should follow the solution treatment. Most of the residue is likely to be quartz sand, glauconite and clay which are to be examined by common methods. In a rather large number of limestones, however, there are residual cherts, not suspected before solution. These may have characteristic sizes, forms and colors. Some are fossils, others oölites from either oölitic limestone or sandstone, still others are "dolocasts," containing cavities from which dolomite rhombs have been dissolved (Fig. 189).

MISCELLANEOUS SEDIMENTARY ROCKS

Bedded Precipitates and "Evaporites" (Other than Limestone)

A. Textures.—Textures appear in the name as qualifiers: massive, crystalline, granular interlocking, granular, coarse to fine, fibrous to earthy, rarely fossiliferous.

B. Minerals.—Minerals are the chief basis of names given.

Rock salt, and potash salts, and other soda salts.¹

Gypsum, rock gypsum.

Anhydrite rock, and other sulphates.

Onyx, siliceous sinter, geyserite, beds of chert.

Phthanite, silicified shale, etc.

Siderite and manganese carbonates (clay ironstone, black band ore).

Greenalite rocks, chamosite rocks, leptoehlorite rocks, and glauconite rocks (Fig. 190).

Ferruginous chert, taconite, etc.

Sulphur, usually mixed with gypsum, calcite etc., or in more common sediments.

Borates, nitrates, etc.

Caliche, both nitrate and carbonate crusts.

*C. Names based on origin have been suggested.*²

Associations of several minerals in precipitates are suggested by the table. The series of 25 or 30 minerals deposited from sea water is most studied and involves the largest volume of rock. The series is (1)

¹ Detailed studies are available in GRABAU, A. W., "Geology of the Non-metallic Mineral Deposits Other than Silicate," McGraw-Hill Book Company, Inc., 1920.

² BERKEY, C. P., New York State Mus. Bull. 251, pp. 105-115, 1922-1924.



FIG. 190.—Photomicrograph of the characteristic granule texture of the Animikian iron-bearing rocks of the Lake Superior region. The granules are believed to have been originally greenalite, but show a variety of alteration products, and internal structures. $\times 20$.



FIG. 191.—A group of calcreous concretions, one to six inches long, from a clay, Koochi ching County, Minnesota.



FIG. 192.—A calcareous septarian concretion eroded so that the cracks and filling show on all sides. Colorado. About one-half natural size.

hematite and CaCO_3 at the base, (2) gypsum and rock salt, with alternations in many places, (3) sulphates and halogen compounds of Ca, Mg, Na, K, in the final residues (page 332).

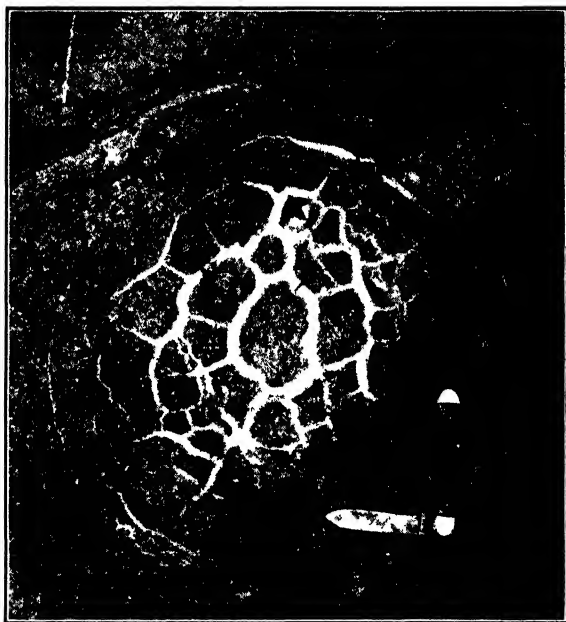


FIG. 193.—A cross section of cherty septarian concretion, Babbitt, Minnesota.

Concretions, Oölites, Nodules, Stalactites, Etc.

Names are based chiefly on the above forms and structures.

A. Textures.—Textures may be noted as qualifiers in the name: botryoidal, pisolitic, with fossils, septarian, geodes, cone-in-cone: (see Figs. 191, 192, and 193).

B. Minerals.—Minerals are also added as qualifiers; calcite and siderite, chert and flint, limonite, goethite and hematite, marcasite and pyrite, quartz, bauxite and clay, manganese oxides, phosphate and barite. Surface crusts of bauxite and even of iron oxides are called “laterites.” Many of the textures suggest colloidal origin (page 334).

Veins

(Smaller veins may be excluded as not rocks)

A. Textures.—Added to the mineral names as qualifiers: massive (near granitoid), banded, comb structure, flamboyant and feathered (Figs. 194 and 195), vugs, druses, inclusions of wall rock, dendrites (Fig. 196).

The criteria listed to distinguish veins from replacements are largely structural or textural.¹

¹ ADAMS, S. F., A microscopic study of vein quartz, *Econ. Geology*, vol. 15, pp. 623–644, 1920.

B. Minerals.—Minerals are these:

Vein quartz.

Vein calcite.

Less commonly fluorite, barite, pyrite, rhodochrosite.



FIG. 194.—Photomicrograph of vein quartz, with a little comb structure and flamboyant structure. Supie Island, Lake Superior. Crossed nicols. $\times 20$. Sand derived from such vein quartz has been mistaken at times, for quartz from metamorphic rocks (Fig. 226).

Replacements

A. Textures.—Pseudomorphs; almost any former structure may be retained; some banding and concretionary forms may be developed; seldom mentioned in naming the rocks.

B. Minerals.—Minerals in replacements are numerous and are the basis of naming the rocks. Common minerals, in addition to the carbonates noted as limestones, are as follows:

Quartz, chalcedony and opal; yield fossil wood, silicified limestone, siliceous oölite, etc.

Phosphorites, largely collophanite, replace calcite limestones, etc. (page 329).

Siderite and hematite replace limestones (forming Clinton ore, part oölitic).

Pyrite and marcasite replace fossils in limestone, and wood in coal, etc. (Fig. 197).

Glauconite replaces mud and shells.

Sericite and chlorite, serpentine, etc., in hydrothermal replacements (see propylite and metamorphic replacements (pages 415 to 420).



FIG. 195.—Photomicrograph of vein quartz with zoned structure and "feathered" structure in the outer zones. Susie Island, Lake Superior. $\times 40$.

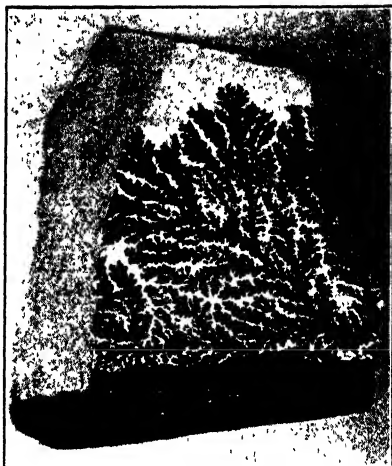


FIG. 196.—Dendrites, branching crystalline forms, largely manganese oxides, that grow commonly in the joint spaces of fine grained uniform rocks. About natural size.



FIG. 197.—Photomicrograph of a polished surface of pyrite from an Illinois coal bed. (After Schwartz.) The organic structures are clearly visible even after the original material is wholly replaced. $\times 25$.

CRITERIA SUGGESTING VEIN FILLING AS DISTINCT FROM REPLACEMENT

(By water solutions, not by ore magmas)

Structure.

1. Sharp contacts between vein material and wall rock.
2. Opposite walls of a vein match at least fairly well.
3. Scalloped or concave surfaces (cave filling rather than true vein).
4. A network of small veinlets.
5. Different orientation of fragments in the veins from wall rock.
6. Fragments from distant, especially the overlying, rocks.

Texture.

1. Crustified banding.
2. Symmetrically lined vugs.
3. Banding around fragments of other material within the vein.
4. Delicately banded chalcedony.
5. Book structure, from a fracture zone.
6. Comb structure.
7. Massive and breccia veins.
8. Stalactites, stalagmites, etc.

Minerals.

1. Quartz with the following characteristics is common: massive vein, flamboyant, zoned, feathered, attached at one end.
 2. The following minerals are fairly common: barite, chalcedony, rhodochrosite, fluorite and calcite.
 3. Negative mineral criteria are the minerals suggesting replacement, especially chlorite, sericite, serpentine.
- Contrast with the criteria of replacement (page 419).

Residual Rocks

(In addition to arkose and clays)

A. Textures.—Textures are commonly earthy, fragmental, porous. Terms added to the mineral names as qualifiers.

B. Minerals.—Minerals are the basis of the rock names.

Bauxite and laterites; see Concretions, (page 294).

Phosphate beds and gravels, residues from phosphatic limestone (page 329).

Iron ores; gossan from sulphides, and residual from silicates. Mesabi ores are from greenalite, Cuban ores are from serpentine, others from various iron compounds.

Manganese ore.

Chert breccia from limestone.

Gypsum from pyritic limestone.

Soils and subsoils (see clays).

Organic Remains, Not Limestones

A. Carbonaceous rocks.

1. **The coal series:**

Peat and turf; unconsolidated vegetation.

Lignite, brown coal; showing woody texture, slaking in the fire and by weathering.

Sub-bituminous coal; black but slaking like lignite.

Bituminous coal, soft coal; black, but does not slake so readily as sub-bituminous coal; burns with a long flame. Often banded with differing textures (Fig. 198).

1. **Bright or glance coal** (clarain and vitrain).

2. **Dull or splint coal** (durain): (a) coking, (b) non-coking.

3. **Waxy or cannel coal** (largely volatile).

4. **Mineral charcoal,** mother of coal (fusain).

Semibituminous coal } gradations.
Semianthracite

Anthracite, hard coal; short blue flame.

(Graphite, a metamorphic residue.)

Bone coal; high in ash.

2. **Sapropel**, black shale, oil shale and oil rock contain bituminous matter (pitch or asphalt) instead of coaly matter, but the distinction is not always sharp.
3. **Oil and asphalt, ozocerite, anthraxolite.**

These rocks are distinguished in the laboratory by tests of (1) moisture, dried out at 105°C.; (2) volatile matter, removed with moisture by heating red hot in a closed vessel; (3) fixed carbon, including all the rest of the combustible matter; and (4) ash, which remains after thorough burning.

The chief distinguishing feature in the coal series is the increasing proportion of "fixed carbon" from peat to graphite. Other distinguishing



FIG. 198.—Bituminous coal, Colorado. The common banded light and dull coal shows woody structure, somewhat deformed, especially in the bright layers. About one-half natural size.

characters are the high moisture content of peat and lignite and the long flame and conchoidal dull surface of cannel.

The United States Geological Survey prefers "subbituminous" to "black lignite" and classes only brown coals as lignite. (See also additional readings on coal.)

Sapropel, bitumen and similar terms commonly refer to liquid or semisolid organic materials, whether or not they constitute a large part of a rock mass. It is said that there are all gradations from oil (or even gas) to coal. Ordinarily, however, coal is solid and largely insoluble in ordinary solvents, yielding gas and tar or bitumen only by destructive distillation. Bituminous coal yields more such tarry matter than anthracite.

The microscopic study of coal requires a special technique and the interpretation of the forms and structures found (Fig. 200) is based largely



FIG. 199.—Lignite from Texas. After exposure to air it checks (network of small irregular cracks) and slakes badly.

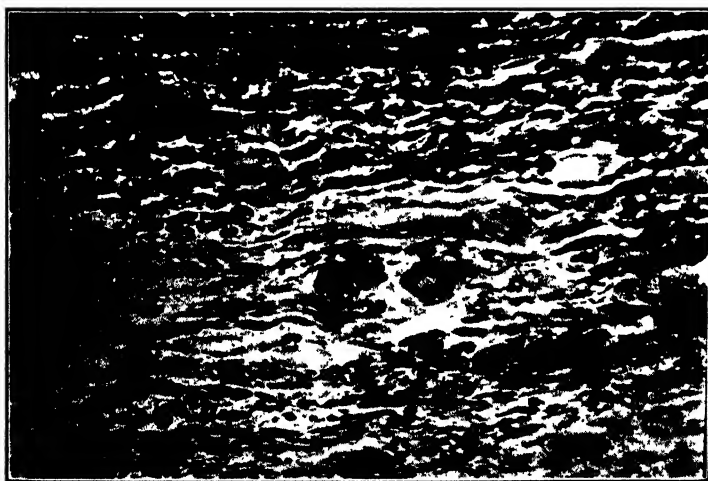


FIG. 200.—Photomicrograph of a thin section of lignite coal from Devonshire, England. Cell structures are clear but somewhat compressed. $\times 70$.

on microscopic botanical work. (See pages 296 and 330, and the list of readings.)

Anthraxolite seems to have less organic structure than coal, and may be altered from asphalt toward graphite.

B. Siliceous earths.

Usually white and fine grained, consisting largely of opal. May mix with deep-sea red clay or with marine shale. Loose earths may become consolidated cherts and flints.

Varieties named from fossil forms (Fig. 201).

Diatomaceous earth (diatomite).

Radiolarian earth.

Sponge spicule earth.

Algal and bacterial (iron formation) cherts.

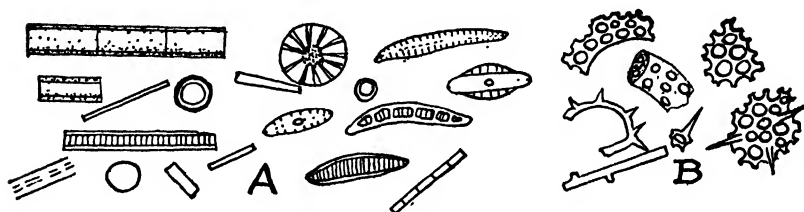


FIG. 201.—Sketches of diatoms (A) and radiola (B) from siliceous earths. $\times 200$.

C. Limonite.

Bog ores are bacterial secretions, with some algae and protozoa (page 327; see also replacements).

Mesabi and Vermilion cherts have ferruginous organic remains.

D. *Manganese oxides* by the same classes of organisms as iron.

E. *Phosphate rock*. (Grades into limestone with no break) (page 329).

Textures of bones, shells, teeth, coprolites, guano, etc.

All are cemented and hence probably somewhat altered.

Minerals, phosphates of Ca, Al, Fe, etc.

F. *Sulphur*. May belong here if due to bacterial action.

MIXED SEDIMENTARY ROCKS

The common minerals quartz, carbonate, and clay may be found mixed in all proportions. The mixtures are usually named from the predominant mineral, and the proper adjective is prefixed to indicate the chief minor constituent. The dividing line might be differently drawn in different places and by different men. A few mixtures have special names, as the "cement rock," a highly argillaceous limestone.

Rarer mixtures, like diatomaceous shale, or shaly coal ("bone coal"), or gypsiferous shale need no special explanation, except that mixtures are common and should be looked for.

Mixtures of sediments with igneous rocks are commonly metamorphic rocks, migmatites; but tuffs are also commonly mixed into shales, and breccia fragments may be mixed into conglomerates.

SUMMARIES OF SEDIMENTS

POROSITY AND SPECIFIC GRAVITY

In the following table after Lane and Barrell, the figures may be a little too high, not allowing enough for porosity; the weight per volume increases as water is eliminated from the minerals.

TABLE XIV.—ESTIMATED SPECIFIC GRAVITIES OF SEDIMENTS

	Gravity	Weight with pores half full, tons per cubic meter
Average igneous rock.....	2.89	2.8
Shale.....	2.69	2.51
Sandstone.....	2.67	2.35
Limestone.....	2.76	2.64
Cenozoic rocks.....	2.67

Somewhat related to the porosity of a sediment is its water-tightness, which governs its usefulness as a cover or base for water-bearing beds, but the size of pores and size of grain are even more important.¹ A clay may have more water in its pores than a gravel when both are saturated. The clay, however, exerts so much more friction as water circulates through it that it is almost water-tight, in contrast with the gravel which allows rapid circulation.

COLOR OF SEDIMENTS

The chief minerals of sediments, kaolinite, quartz and calcite are colorless in their pure state, and some considerable bodies of clay, sandstone and limestone are white. The colors of others are usually due to a relatively small quantity of some pigment coating or staining the white grains or forming a cement between them. The commonest pigments are minerals of iron and carbonaceous matter, each of which has several colors. The chief iron colors are red, brown and yellow, rarely green and black; the organic colors are black, green and brown, rarely red or yellow. If both iron and organic colors are present the organic color seems to dominate.

Minor coloring matters include black and brown manganese oxides and green glauconite and chlorite. Fine-grained pyrite and other iron sulphides furnish black colors, and are said to be present in the "blue mud" on the margins of continental shelves.

A red limestone is to be interpreted as weathered, either as exposed or before. Eolian and continental deposits are found to be most variegated in alternating bands of bright colors.

¹ MEINZER, O. E., The Carlsbad irrigation project, New Mexico, with respect to water-tightness, U. S. Geol. Survey, Water Supply Paper 580 A, pp. 16-28, 1927.

The conditions under which a shale or shaly sandstone may be deposited so as to form "red beds" are less certain; there seem to be several possibilities. See the further readings on red beds listed.

COMPOSITION

TABLE XV.—ANALYSES OF CERTAIN SEDIMENTS AND SEDIMENTARY MINERALS
(After Clarke, Leith and Mead, and others)

	Average shale	Kaolin-ite	Average sand-stone	Average lime-stone	Lime-stone for build-ings	Calcite	Dolo-mite
SiO ₂	58 90	46 5	78.64	5 20	14 09		
Al ₂ O ₃	15 63	39.5	4.77	0.81	1.75		
Fe ₂ O ₃	4 07	1.08	0.54	0.77		
FeO	2.48	0.30				
MgO	2 47	1.17	7 92	4.49	21.9
CaO	3 15	5.51	42.74	40 60	56.04	30.4
Na ₂ O	1.32	0 45	0.05	0 62		
K ₂ O	3 28	1.32	0.33	0.58		
H ₂ O +	3.72	14.0	1.33	0.56	0.88		
TiO ₂	0 66	0 25	0.06	0.08		
P ₂ O ₅	0.17	..	0.08	0 04	0 42		
CO ₂	2 67	5 03	41 70	35 58	43.96	47.7
Miscellaneous ..	1.48	0.07	0.05	0 48		
	100.00	100 00	100 00	100 00	100.34	100 00	100.00
Estimated minerals:							
Quartz	30	...	70	4			
Kaolinite	12	100	8	1			
Mica	18	...					
Dolomite	3	...	3	36	..		100
Calcite	5	...	7	56	..	100	
Feldspar	12	...	8	2			
Chlorite, etc ..	8	...	1				
Zeolites	4						
Fe oxides, etc ..	5						
Miscellaneous ..	3						

HISTORICAL GEOLOGY OF SEDIMENTS

In general, sediments of the three chief clans occur throughout the geologic column, but certain minor peculiarities may be noted. Daly has suggested that the pre-Cambrian organisms may have secreted less

AVERAGE RATIO OF CALCIUM TO MAGNESIUM IN LIMESTONES OF DIFFERENT PERIODS

Period	Number of analyses	Ratio of CaCO ₃ to MgCO ₃	Ratio of Ca to Mg
All pre-Devonian	392	2.39:1	3 35:1
Devonian	106	4.49:1	6 29:1
Carboniferous	238	8.89:1	12.45:1
Cretaceous	77	40.23:1	56.32:1
Tertiary	26	37.92:1	53 09:1
Quaternary and recent	26	25.00:1	35.00:1
	865		

carbonate than those of later periods,¹ also that the ratio of CaCO_3 to MgCO_3 varies through geologic time.²

Iron formations and graywacke sandstones seem to be more abundant in pre-Cambrian than in later rocks.

Less garnet is to be expected in old sands, because garnet was formed by metamorphism at depth and was not exposed until after erosion.

MINERAL ASSOCIATION IN SEDIMENTS

The rules of association in sediments are very different from those in igneous rocks, in which minerals melted together react chemically. In sediments the bases of association are (1) hardness, (2) stability under weathering conditions and (3) possibly the place or conditions of deposition. Clays being soft are mechanically broken up to finer grains and largely sorted away from coarser sands but are associated with iron oxides which are equally fine grained. The hard, sandy minerals that resist weather, like quartz, tourmaline, zircon, and garnet, are commonly deposited with each other. The deep-sea red clays are not expected with shallow sediments such as conglomerates. Red sands and gypsum are not expected in association with coal or such sediments as imply humid climates.

The characteristic sequence in a geosyncline is sandstone grading upward through shale into limestone. It is much less common to find sandstone followed directly by limestone or alternating with it. The regularity of this sequence, however, is only a general rule with many exceptions.

ECONOMIC MATERIALS FROM SEDIMENTARY ROCKS

Building materials:

Sandstones.

These are especially resistant to fire.

Brownstone (formerly very popular).

Concrete aggregate.

Flagstone and paving blocks.

Limestone.

Well-jointed blocks.

Travertine.

Dolomite.

Other varieties.

Concrete aggregate.

Lime for mortar and plaster.

Cement.

Sand and gravel, for mortar and concrete.

Clay, for brick, tile, and cement.

Gypsum, alabaster, and hard wall plaster.

Onyx, ornamental stone.

Diatomaceous earth, for soundproof walls.

Abrasives:

Flint for ball mills.

Diatomaceous earth, garnet sand.

Sand, grindstones.

¹ DALY, R. A., The pre-Cambrian ocean, Eleventh Geol. Cong., p. 503, 1910.

² DALY, R. A., Evolution of the limestones, Geol. Soc. America Bull., vol. 20, p. 165, 1909.

Novaculite, geyserite.

Reservoir for oil and artesian water, sandstone, dolomite, and other porous rocks.

Impervious containers for oil and water, clay.

Soil is fundamentally a residual sediment.

Salt, from rock salt.

Refractories, sand for ganister, dolomite (and magnesite) for furnace lining, fire clay.

Miscellaneous. CO₂, lime salts, and soda from limestones; pigment from barite; red fire from celestite; potash from salt beds; sulphur and sulphuric acid from sulphurous beds and pyrite veins; fertilizer from phosphates; absorbents from siliceous earths; gas, tar, coal, and coke from coal; glass from sand and limestone; heat insulators from diatomaceous earth.

Ores that are sedimentary.

Placers of gold, cassiterite, magnetite, monazite and diamonds

Iron, manganese and aluminum ores and protores are commonly chemical or organic precipitates or other rocks, enriched secondarily by weathering.

Many veins from solutions in water carry metals of value.

Replacements of limestone and other sediments include hematite and the ores of lead, zinc, and silver.

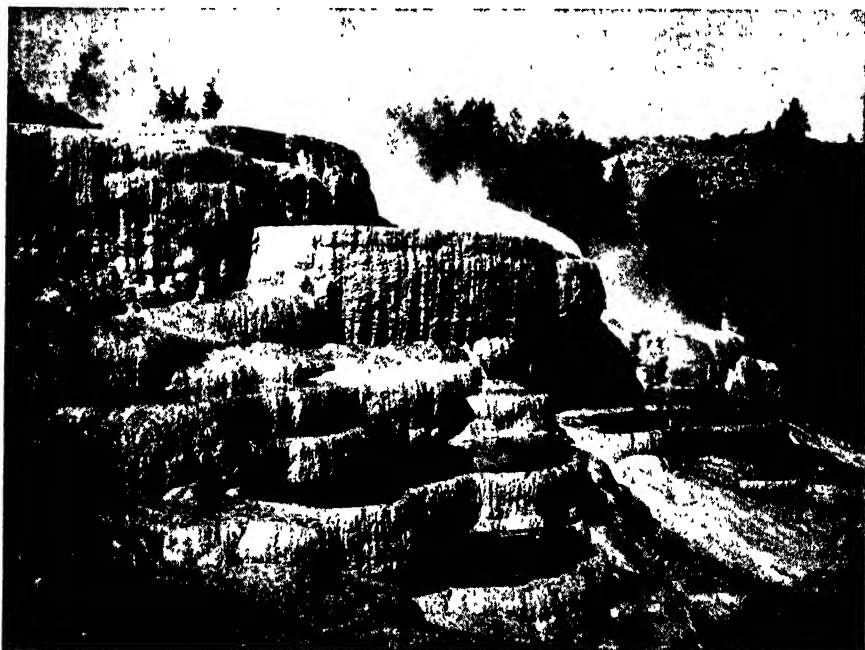


FIG. 202.—The terraces of travertine in Yellowstone Park. (Photograph by J. E. Haynes, St. Paul.) Cleopatra Terrace. In these pools rocks of aqueous origin can be seen in the process of formation.

Sedimentary rocks that are favored hosts of ores.

Limestone, the favored host of lead, zinc, copper, and silver.

Sandstone

Carnotite is rare except as cement of a sand.

Copper ores may cement a sand or clay.

Lead-silver ore commonly replaces sandstone.

Conglomerate

Copper occurs in and replacing Calumet conglomerate.

Gold occurs introduced in Rand conglomerate.

Clay

Unfavorable to introduced ores but may control or limit the deposition.

Residual ores of manganese and iron occur in residual clay.

PART V

THE PETROLOGY OF SEDIMENTARY ROCKS

SOURCES OF MATERIALS

The chief sources of sediments may be grouped as (1) débris of earlier rocks broken up mechanically; (2) residues from leaching of earlier rocks; (3) solutions formed by leaching; (4) solutions from igneous emanations. Minor sources include (5) cosmic dust, which forms part of the deep-sea red clay; (6) the atmosphere supplies snow and some of the carbonaceous matter of organisms.

Most of these sources involve the process of weathering. The igneous emanations contributed to sediments are the same as those already described as igneous effects (page 216). The minor sediments of the deep sea and the snow from the atmosphere need no comment here. The hydrocarbon compounds produced from the carbonic acid of the atmosphere are well understood to depend on plant life.

WEATHERING

Weathering is the alteration of rocks near the earth's surface, chiefly by air, water, and water solutions. It is partly a mechanical disintegration and partly a breaking down into new minerals and solutions. The mechanical process may produce a fragmental rock almost without mineral change or transportation, but in most cases some of the minerals are at least partly decomposed. It is probable that there is no constituent of rocks that is not chemically attacked by some of the reagents involved in weathering at or near the surface. Many minerals are attacked so slowly as to be common in weathered residues. Chief among these is quartz. The most stable decomposition products are hydrous compounds of aluminum and iron.

Mechanical Processes.—Mechanical processes include action by air, water and ice, action by organisms such as growing roots of plants, and volume changes resulting from either daily or seasonal temperature changes, or from chemical action. Rocks are worn or broken by air and water chiefly through sand and gravel carried against the solid rock by winds, streams and waves. These are the cutting tools by which air and water carve solid rock. Similarly, but perhaps in a more striking way, ice carves rocks by means of the boulders frozen in at the borders of the moving ice.

The change of volume by reason of daily or seasonal change is probably small in comparison with the change at times of forest fires, though the

alternations are more numerous and may have some effect in a period of years. Chips are forced off from rocks by the expansion of water freezing in crevices. The spalling produced by a single forest fire is probably

TABLE XVI.—ALTERATION OF COMMON MINERALS
(Modified from Holmes)

Deep-seated minerals	Hydrothermal alteration products	Weathering products		Minerals deposited from solution
		Secondary minerals	Removed in solution ¹	
Quartz	SiO ₂	Quartz, opal chalcedony
Feldspars Orthoclase Plagioclase	Sericite Kaolinite? Zeolites Epidote Paragonite Kaolinite?	Kaolinite Other clay minerals Bauxite	K ₂ CO ₃ SiO ₂ Na ₂ CO ₃ CaH ₂ (CO ₃) ₂	Salts Calcite Gypsum Glauconite, etc.
Feldspathoid	Zeolites	Kaolinite Bauxite	Na ₂ CO ₃ K ₂ CO ₃ SiO ₂	Salts
Muscovite	Sericite	Hydromicas clay, etc.	K ₂ CO ₃ SiO ₂	Salts
Biotite	Chlorite	Limonite, clay, etc.	(Mg, Fe)H ₂ (CO ₃) ₂ K ₂ CO ₃ SiO ₂	Siderite Glauconite, etc.
Augite	Hornblende Biotite Chlorite	Limonite Clay minerals	SiO ₂ CaH ₂ (CO ₃) ₂ MgH ₂ (CO ₃) ₂ FeH ₂ (CO ₃) ₂	Calcite Dolomite Siderite
Hornblende	Biotite Chlorite			
Enstatite and olivine	Serpentine Talc Iron oxides	Limonite Hematite	SiO ₂ MgH ₂ (CO ₃) ₂ FeH ₂ (CO ₃) ₂	Siderite Magnesite Limonite
Magnetite	Pyrite	Limonite	FeH ₂ (CO ₃) ₂	Limonite
Pyrite	Pyrite	Limonite	FeSO ₄ H ₂ SO ₄	Limonite Gypsum
Ilmenite	Leucoxene rutile	Leucoxene Limonite, etc.	FeH ₂ (CO ₃) ₂	Limonite
Garnet	Chlorite	Limonite Clay minerals	Bicarbonates	Limonite Carbonates
Apatite	Ca ₃ (PO ₄) ₂	Phosphorite, etc.

¹ SiO₂ dissolves as alkali silicate and colloidal silicic acid; CO₂ is supplied mostly by the atmosphere. For chemical effects see LEITH and MEAD, "Metamorphic Geology"; STEIDTMANN, *Econ. Geology*, vol. 3, pp. 381-409, 1908.

comparable to the effects of sun and frost for 1,000 years, though the great effect of a fire may be partly a result of previous loosening of grains by sun, frost and other agencies. Deep weathering to residual boulder-like masses cannot be considered a result of forest fires. Commonly spheroidal structures result (Fig. 203). This is not a sign of altitude or desert conditions, but rather of a moist situation, where the hydration of feldspars and other minerals causes exfoliation. The effects of temperature change may be subordinate to the mechanical effects of hydration.

Gravel is produced chiefly subaerially,¹ though the pressure of ice may break some weak rocks, and even wave action may develop a



FIG. 203.—Spheroidally weathering diabase. Duluth, Minnesota.

force of a kilogram per square centimeter and break off fragments of shore cliffs. Norton has discussed the criteria of origin of sedimentary breccias (page 276).

Most erosive action is exerted on the surfaces of land masses of wide area rather than on coast lines, so that the source of a sediment is not the nearest cliff of rock exposed, except in local instances. The sediments from river and rain erosion are probably fifty to one hundred times more abundant than those from marine erosion of shores. Sediments transported for great distances usually receive minerals from a variety of rock sources.

Probably the most abundant product of mechanical disintegration without much decomposition and with relatively little transportation is

¹ BARRELL, JOSEPH, Marine and terrestrial conglomerates, *Geol. Soc. America Bull.*, vol. 36, p. 289, 1925.

arkose, formed from granite and gneiss. Graywacke is a somewhat equivalent product from darker original rocks. Detailed studies indicate that most arkose develops in a rigorous climate, but it may be either desert, or semiarid, or at high altitudes or otherwise. Most arkoses overlie granite; graywackes commonly overlie less-silicic rocks. (See the further readings listed.)

Chemical Processes.—The solutions formed by the leaching of rocks pass into ground water and in large part into the rivers, so that analyses of the river waters show the main results of leaching in a broad way. In mountainous regions the dissolved metals are Ca, Na, and Mg, and the acid radicles are CO_3 , SO_4 , Cl, and SiO_2 in the order given. On the plains Na and SO_4 become more important, either by precipitation of early solutions or by additions from the rocks of the plains. Note that K and SiO_2 are not largely carried in solution and that the CO_2 is largely from the atmosphere.

AVERAGE COMPOSITION OF SOLIDS IN SOLUTION IN RIVER WATERS
(After Clarke)

	Per cent		Per cent
CO_3	35.15	Na	5.79
SO_4	12.14	K	2.12
Cl	5.68	$(\text{Fe}, \text{Al})_2\text{O}_3$	2.75
NO_3	0.90	SiO_2	11.67
Ca	20.39		
Mg	3.41		100.00

Table XVI shows the source of the main bulk of our sediments, for it includes the products of weathering of the most abundant minerals in the earth's crust. Nevertheless it may be deceptive in one respect, for a considerable amount of sediment is the result of mechanical breaking up of minerals that resist chemical alteration. The chief of these is quartz, which forms sand, but there is much feldspar and a considerable list of accessory and metamorphic minerals that are also slow to alter (Fig. 181). The minerals of sand depend on (1) source, (2) length of transport, and (3) the agents of decomposition.

Solvents for Particular Constituents and Unusual Precipitates.—Most chemical attack upon minerals is attributed to carbonated waters, but organisms and organic acids may greatly modify the solutions. It is clear also to anyone with experience that conditions of attack must have differed in different cases, because the results are so unlike. In a certain environment silica may be leached from iron, and in another iron oxides may be leached away from silica. Orthoclase may be more attacked than plagioclase in some rocks, whereas others, having been subjected to a

different attack, show the plagioclase most altered. The reagents in solution must include a considerable variety in different cases.

Silica is leached especially by alkalies and by $\text{MgH}_2(\text{CO}_3)_2$ solutions.

Alumina migrates only a little but is soluble as sulphate and in organic solvents. Bacteria may play an important part in the migration of alumina.¹ It has been suggested further that the kind of solution may determine whether aluminous minerals are altered to kaolin or to some other hydrous compound. Tropical weathering gives laterite, containing bauxite; ordinary weathering gives kaolinite (dickite from hydrothermal action) and in many cases an even more-siliceous clay mineral. Experiments indicate that bauxite is residual, if alkali carbonate solutions leach the rock.

Iron oxides vary but in general are leached by humic acids and as ferrous bicarbonate, sulphate, chloride, etc. Ferric oxides are less soluble leaving gossans.

Calcium, magnesium, sodium and potassium are soluble as sulphates and carbonates or bicarbonates.

Phosphates are soluble in humic and carbonic acid solutions.

These reagents are complicated by development of colloidal solutions, especially of silica but also of some metallic compounds. Some constituents may be "peptized" (kept in solution) by organic colloids.

Certain unusual precipitates must have unusual supplies of solutions.

1. Great chert formations probably get their silica from igneous emanations or other hot solutions or from diatomaceous skeletons secreted in sea water.

2. Great iron formations may result from a peculiar weathering or from igneous emanation.

3. Great salt beds derive sodium from the weathering of igneous rocks containing soda and derive chlorine from volcanic emanations. Thick beds probably result from prolonged supplies of salt sea water (page 332).

4. Great gypsum beds may form similarly from supplies of sea water.

5. Great phosphate deposits are probably all enriched (1) by reaction of sea water on slightly phosphatic shells, (2) by surface leaching of calcite from a lean phosphate rock or (3) by secondary enrichment by descending water (see page 340).

6. Borates seem to develop from igneous emanations concentrated by the evaporation of ground water.

7. Nitrates may result from organic fixation of atmospheric nitrogen or electrical reactions on the air; possibly also from igneous action.

8. The potash of shales is a problem, because many shales have notable amounts of potash in some unknown form. Apparently it is in two forms, one more soluble and more available as plant food than the other. Now the weathering of feldspar is known in a number of cases to remove potash very completely leaving kaolinite. Before the leached potash reaches the sea, however, it seems to react upon or be absorbed by the clays, so that no great amount of potash concentrates in the sea water as sodium does. Probably much of the potash as first precipitated in the mud, is colloidal, but there may be zeolites, sericite, glauconite, and other minerals.

¹ THIEL, G. A., The enrichment of bauxite deposits through the activity of micro-organisms, *Econ. Geology*, vol. 22, pp. 480-493, 1927.

Conditions and Results of Weathering.—It will be noted from the foregoing discussions that the conditions that favor weathering are varied. Rocks are most weathered above the water table, in warm moist climates, where the topography favors considerable circulation of water. The rocks most rapidly weathered are the porous, fine-grained rocks with soluble minerals, notably the carbonates, basic silicates, pyrite, or volcanic ash. The porosity may be primary, as in a fragmental rock or in an amygdaloid, or secondary, in sheared metamorphic rocks.

The results of weathering normally include a series of zones from the surface down to the fresh rock at depth.¹ The order is

4. Ferruginous crusts.
3. Leached zone.
2. Decomposed zone.
1. Fresh rock.

It is also noteworthy that although complete weathering results in rocks of relatively simple mineralogic nature, partial weathering may produce a mixture having all the complexity of the original, intermediate, and final stages—among the most complex rocks known. For this reason two sets of criteria are here tabulated in connection with weathering; (1) for ordinary weathering, (2) for the original rocks from which a weathered rock has been formed.

CRITERIA OF WEATHERING

Mineralogic.

Minerals usually formed by weathering or related processes. (Hydrothermal minerals (pages 424 to 427) are only remotely suggestive.)

Hematite.	Leucoxene.
Limonite (various forms).	Bauxite.
Kaolinite (and other clay minerals).	Psilomelane.
Chalcedony.	Pyrolusite.
Opal.	Alunite.
Native metals.	Carbonates.

Metallic oxides, carbonates, chlorides, sulphates, hydrous silicates, etc. Partial weathering makes very complex rocks, but complete weathering makes them mineralogically simple.

Textural and structural.

Textures and structures most common in weathered rocks.

Solution cavities, spongy, cellular, honeycombed, mamillary, concretionary, stalactitic, oolitic and pisolitic; slumping due to leaching.

Imbricating blades of quartz, which join at angles representing the cleavage of calcite.

Negative crystals.

Development of cracks along cleavage planes.

Pseudomorphs.

Dreikanter and desert varnish.

Negative.

The presence of easily weathered minerals, such as pyrite, ordinary silicates, carbonates of Fe, Mg, and Ca in an unaltered state, suggests that no weathering has taken place.

¹ This zoning has been studied in detail by HARRASSOWITZ, *Fortschritte Geol. u. Palaeont.*, Berlin, 1926. See also MEAD, W. J., The bauxite deposits of Arkansas, *Econ. Geology*, vol. 10, pp. 28-54, 1915.

Color.

The presence of variegated colors such as red, pink, yellow, and brown indicates weathering.

Iron stains, gossans, and blooms are signs of weathering.

Field and general.

Concentrations of minerals resistant to weathering indicate that other minerals have been removed by weathering. Placer deposits are examples of this concentration. Common minerals resistant to weathering are gold, platinum, quartz, garnet, magnetite, chromite, hematite, cassiterite, monazite, rutile, zircon, and galena.

Weathered rocks may merge into original, or have residual cores unaltered.

Weathered rocks show a relation to surface and ground waters and sometimes to structural features.

Secondarily enriched ores show that weathering has taken place, usually above them.

Completely weathered zones in tropical climates may develop an upper crust of iron oxide or bauxite over the leached rock, in many places concretionary (Lateritic.)

Criteria for mechanical sedimentation (page 313) are suggestive.

Exfoliation.

CRITERIA IN WEATHERED RESIDUES, SUGGESTING NATURE OF ORIGINAL ROCK**Textures.**

1. Angular quartz in clay with granitoid texture—granite or granodiorite
2. Rounded sand and pebbles—mechanical sediment.
3. Parallelism of platy chlorite, etc.—schist or gneiss.
4. Diabasic, porphyritic, amygdaloidal and other textures may be recognized in some residues.

In hand specimen.

1. Talus and breccia and some residual soils contain fresh remnants of parent rock.

2. Arkose—from granite.

3. High percentage of SiO_2 —from silicic rocks.

4. Clay with quartz grains and silicic feldspar fragments, etc.—from a silicic igneous rock.

5. Dark-green ferromagnesian sand—from a basic rock.

6. A dark or green clay containing abundant biotite, chlorite, talc, serpentine, magnetite, hornblende fragments, etc.—from an original basic rock, igneous or metamorphic.

7. Laterites; if low in iron oxide and TiO_2 and high in Al_2O_3 —from silicic rocks; if high in iron oxide and titania and low in alumina—from a basic rock. Both may be from metamorphic rocks.

8. Gossan—from a sulphide deposit, or less often from siderite.

9. Red colors. Most red from basalt, gabbro, etc. Moderate; terra rossa, a fine red clay, in which abundant iron gives concentric concretions coated with limonite—from limestone. Light; most sediments. Whitest; may be gray or yellow—granite and rhyolite.

10. Clay or soil high in phosphate, manganese, barite, etc.—from limestone.

11. Clay with flints or cherts—from limestone beds.

12. A great variety of residual minerals suggests igneous or metamorphic original rock.

13. Bentonite, which swells greatly in water—from volcanic ash.

14. Abundant mica usually from a mica schist.

15. Kaolinite—from feldspar and other aluminous silicates.

Bauxite—from nephelite and other aluminous silicates.

Limonite—from many ferromagnesian minerals, serpentine, peridotite, iron formation.

16. Clay of sedimentary origin can commonly be distinguished from residual clay by its greater plasticity and swelling power; not so great as bentonite.

In washed residues of coarse and heavy minerals.

a. Quartz grains containing acicular inclusions suggest granite.

b. Quartz grains containing inclusions of chlorite, mica, rutile, magnetite indicate a schist original.

- c. Chromite, diamond, platinum—from peridotite.
- d. Cassiterite and zircon in euhedral grains, titanite and topaz—from granite.
- e. Ilmenite and leucoxene—from gabbro.
- f. Corundum—from nephelite, syenite, peridotite, schists.
- g. Sillimanite, andalusite, kyanite, staurolite, garnet—from schists of sedimentary origin.
- h. Tourmaline, beryl, cassiterite—from pegmatites.
- i. Chlorite—from biotite, hornblende, etc.
- j. Talc and serpentine—from olivine, pyroxene and altered limestone
- k. Mica persists especially well in mica schist areas.

In the field.

Topography: softly undulating—massive rocks.

sink holes—limestone.

ridges suggest dikes or veins or folded sediments.

Water table likely to be higher in granite than in sediments.

Gradation from fresh to weathered rock.

Plants; local observations best.

Summary of Weathering.—In summary of the weathering of our abundant rocks, granites yield arkose and sandy clay soil, and by transportation ordinary sands and clays; gabbros, basalts, and andesites yield graywackes and the more ferruginous clays and soils. The abundant feldspars dominate the chemical change, and their product kaolinite forms a large part of the residue. Darker primary silicates, if they contain alumina, also form clays, but usually they are highly impure with iron oxides, and there are more prominent solutions of the metals. Extreme weathering especially under tropical conditions leaves bauxite and iron oxides. The volume of the products is considerably greater than that of the igneous source rock, partly because of added water and carbonic acid, and partly because sediments have a much higher average porosity.

Sedimentary rocks, chiefly clays and sandstones, are more stable under conditions of weathering, but limestones weather to residual clays.

Textures due to weathering are usually massive to irregular or earthy, but some oölites are formed especially in bauxite.

Chemically the changes are chiefly oxidation, hydration and carbonation through the agency of air and water solution. Rarely also HCl, H₂SO₄, and organic acids are active. Additional points in regard to weathering are included in the readings listed.

TRANSPORTATION OF SEDIMENTS

Sediments are transported chiefly by mechanical suspension in some moving agent but partly also by moving solutions, by simple slumping and, by sliding on a steep slope.

MECHANICAL TRANSPORTATION

Aside from the relatively short movements involved in the slumping of residual deposits and in land slides, rock flows, and the falling of talus or

scree, the agents of transportation are wind, ice and water. The criteria by which to determine that a rock is a mechanical sediment are listed here.

CRITERIA SUGGESTING MECHANICAL TRANSPORTATION

Rocks. Sandstones, shales, etc., usually of simple mineral composition.

Characteristic minerals.

Usually much quartz, feldspar or clay.

Carbonates, as calcite, aragonite, limestone, dolomite less common.

Presence of residual minerals, as

Magnetite.

Ilmenite.

Garnet.

Zircon.

Rutile.

Tourmaline.

Diamond

Gold.

Platinum.

Cassiterite.

Spinel.

Corundum.

Cannel coals and carbonaceous matter.

Usually an absence of more easily altered minerals.

A variety of minerals of different sources.

Textures.

Fragmental, more or less cemented.

Grains are rounded or corroded.

Grains are sorted according to size.

Contains fossils.

Bedding planes, laminations, banding, stratification, layers, laminae.

Usually porous.

Presence of organic material.

Presence of concretions (not including those in limestones).

Component grains of the aggregate do not show interlocked outlines.

Shards.

Striated pebbles.

Structures.

Bedding.

Cross-bedding and imbricate structure.

Ripple marks and rill marks, rain and hail marks.

Mud cracks.

Footprints.

Absence of any noticeable or conspicuous structural features.

Aggregate may consist of grains that show different geologic histories.

Sorting of grains of different sizes.

Color. Red, brown, white, blue black.

Wind carries some fine particles over long distances. It is believed that the finest dust and volcanic ash may be borne all around the earth before settling. Storms carry coarser dust several hundred miles, and sand grains may go several rods at a single leap. Gravel in grains up to 8 mm. in diameter is seldom lifted by wind and then moved only a few feet at a time. There is thus a good deal of sorting in the process of picking up the grains transported by wind, but nearly any mineral may be carried; feldspars are common in addition to quartz. Minerals are sorted by wind according to specific gravity also, where the wind blows up a slope.¹ The grains transported by wind, even the smaller sizes, become well rounded, and sand grains are "frosted" on the surface. Pebbles over which sand is blown become faceted into "dreikanter." The transportation develops wind ripples and dune structures in sandy beds.

¹ BARRETT, W. H., *Geol. Mag.*, vol. 67, pp. 159-161, 1930.

Transportation by ice involves little or no sorting. Large grains and small are picked up and carried together. The grains are not rounded during transportation, though rounded grains may be carried with others without much change. Pebbles and boulders become somewhat flattened on one or more sides and scratched or grooved ("striated"). The bed-rock is also grooved and striated.

Transportation by water has been most studied, and most sedimentation involves this transportation. The law of the transporting power of water commonly given in the elementary texts is that the size of particles carried varies as the sixth power of the velocity; $S = K.V^6$. This holds only under practically uniform conditions and is ordinarily modified by the form and density of grains, by the depth, width and straightness of the channel and by several other factors. An easy value to remember as a basis for calculation is that water flowing one foot per second moves quartz grains about half a millimeter in diameter.

TABLE XVII.—APPROXIMATE CURRENT VELOCITY FOR DÉBRIS MOVEMENT IN SHALLOW WATER
(After Bucher)

	Mean diameter, millimeters	Velocity, meters per second
Clay	0 08
Sand.....	0 5	.28
Coarse sand.	1 0	.40
Fine gravel	6.0	.70
Gravel.....	54.0	1 62
Boulders.....	750 0	11 69

It is to be noted first of all that there are several *modes of transportation* by streams. (1) Rolling is easily understood. (2) "Saltation" consists of short jumps that result from the force of flowing water giving a particle an upward momentum as it rolls up over an irregularity in the bed. (3) The finer particles of clay take longer jumps than sand and practically stay in suspension, some being so fine as to constitute colloidal solution. (4) Water may so lubricate some grains in the bed of a stream or delta as to make possible a "collective movement" of the wet mass. Subaqueous "gliding" is the best explanation of some complex structures in shaly beds (see page 341).

Water in mountain torrents can transport some surprisingly large boulders down steep slopes, and many large boulders have probably been carried by ice. Some sorting occurs as material is picked up and carried in streams of varying velocity, a stream with any particular velocity carrying certain sizes and leaving behind the coarser, according to the law of transporting power. Grain size varies with distance of transporta-

tion, though the original size, the agents of chemical attack, and agent of transportation modify the effect on grain size. Pebbles may be rolled 500 to 2,000 miles before being entirely worn down to sand.

Reworking, or repeated transportation, and probably also long-distance transportation tend to concentrate grains of one size and one specific gravity in a formation. A sand 99.9 per cent silica is probably derived from an earlier deposit of sand. The mineralogic effects of sorting are chiefly that clay minerals and iron oxides are concentrated in the clays, and quartz is carried largely as fine sand. It is said also that the variety of abundant minerals in a formation varies inversely as the distance transported.¹ Streams characteristically round the pebbles and sands transported if the size is no smaller than about 0.05 mm. Smaller grains are not well rounded in water. The grains of water-worn sand are clear or polished (Fig. 180). It is believed that transportation by water rounds the sand grains better than the same amount of transportation by air, but both agents of transportation round the sand grains so slowly that a well-rounded sand suggests that it may be reworked—twice transported.

The sedimentary structures developed by transportation are not distinct from those treated later as structures of deposition. See the further readings on transportation.

Summary of Form of Grain and Transportation.—The chief bulk of fragmental sediment is clay with grains too fine for satisfactory studies of form. Next come the sand grains, the forms of which have been considerably studied but with results that are very unreliable as criteria for determining the origin of the sand deposit. The coarser pebbles are studied with better results, but conclusions even from these cannot be stated too positively. The following points, therefore, are more suggestive than determinative.

Although rounding is not necessarily a result of abrasion, and water will dissolve some sharp corners from grains, it is nevertheless safe to conclude that most well-rounded grains have been abraded. The factors determining form are (1) jointing, bedding, cleavage, crystal form, or other antecedent structures in the pebble or grain; (2) solution; (3) rolling by streams; (4) glaciation; (5) wind abrasion; (6) rubbing or shoving along the bottom, as waves disturb a beach gravel (Fig. 204).

Sharp angular grains indicate a short transport or glacial action. Many glacial pebbles are striated. Rounded grains may be of various shapes, but an accumulation of disc-like pebbles suggests beach action. If a sand (not calcareous) has more than 50 per cent of the grains well rounded it suggests wind action, but less than 50 per cent gives no indication as to origin. Water can round sand grains as small as 0.05 mm., but wind may round grains of 0.03 mm. or smaller. Wind-blown sand has

¹ BOSWELL, Liverpool Geol. Soc. Trans., vol. 14, pp. 2-3, 1923-1924.

frosted grains, whereas water-worn grains are polished. A few wind-blown grains in a sand do not indicate that all the rest were wind blown. Coated grains of sand indicate weathering. Characteristic dreikanter indicate wind erosion; they are said to have faces better polished and angles sharper than glacial pebbles.

CRITERIA OF AGENTS AND DISTANCE OF TRANSPORTATION OF SEDIMENTS

Ice.—The evidence consists of striated faceted boulders, absence of sorting and uncemented heterogeneous mixtures of angular pieces. There may be grains of any size and shape without kaolinization or structure, but rocks are of large extent. Many have large boulders in great deposits.

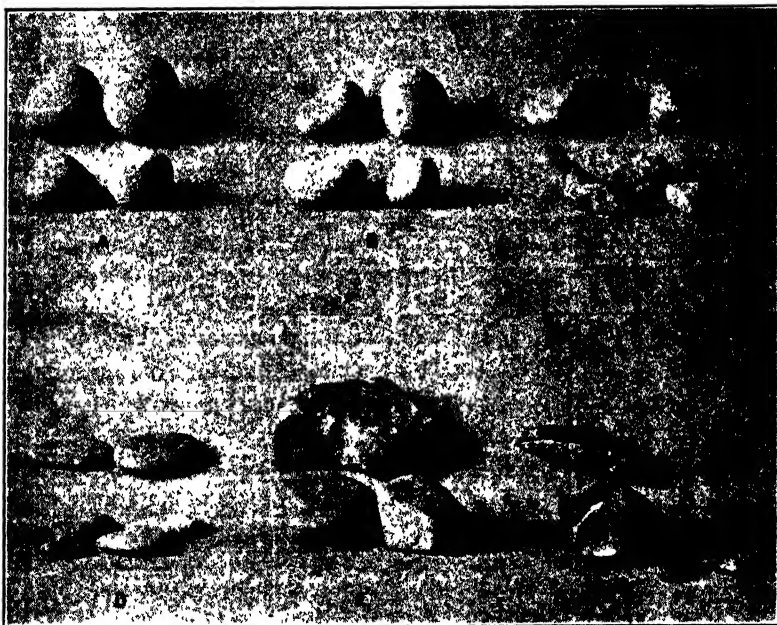


FIG. 204.—The shapes of pebbles. *A*, rounded; *B*, sub-rounded; *C*, angular; *D*, Disc-like pebbles; *E*, faceted and striated glacial pebbles; *F*, dreikanter, faceted by wind-blown sand. About one-fourth natural size.

Water.—Fairly uniform in both composition and grain size, with polished surfaces and rounded corners. Deposits may be cemented, and some structure will always be present, such as bedding and cross-bedding. The latter will be tangential to the enclosing beds, at least at the base, which is usually flat, but concave from current action. Large boulders only locally.

Air.—Same as water but the grain size is smaller (not more than 5 mm. in any extensive beds), more rounded, frosted rather than polished, usually uncemented, and cross-bedded on a curved base rather than uniform, tangential. Grains may be rounded even if below 0.03 mm. (Loess is extensive but has fine angular grains.) All deposits well sorted. Dreikanter are associated.

Distance.—If transported by water, pebbles have their sharpest edges rounded to a radius about one-tenth the average radius in a mile or two. Field studies indi-

cate that most round stones are worn down to sand within 1,500 miles, many within 500 miles.

TRANSPORTATION IN SOLUTION

Table XVI (page 306) shows the main solutions that result from weathering, but it is to be noted that all other substances may be carried, usually in minor quantities, and rarely concentrated except by precipitation. Most silica seems to be carried in colloidal form, as a number of other substances may be. The abundant metals are more commonly transported in solution as sulphates and carbonates or bicarbonates. Chlorides may be added, but rarely in large quantities, except by volcanoes or by human agencies.

These solutions, carrying their dissolved load, flow under hydrodynamic conditions. They circulate in ground-water channels and constitute streams descending to the sea or inland lakes.

Certain features of their environment may prove unfavorable to continued transportation in solution. (1) Oxygen is likely to precipitate iron from surface waters. (2) Clays and other colloids absorb potash from solution. (3) Mingling solutions may react to precipitate salts. (4) Agitation, evaporation, or changes of temperature and pressure may remove some salts from solution. (5) Electrolytes tend to coagulate colloidal solutions.

Various surface crusts deposited from solution may have been transported under the control of capillarity rather than gravity, namely, the nitrates, carbonates, borates, caliche, and possibly some bauxite and iron crusts.

DEPOSITION OF SEDIMENTS

MECHANICAL DEPOSITION

Places of Deposition.—A good deal of sediment may be transported by streams or winds in short stages alternating with temporary deposition on land surfaces. Such continental deposits, although they may be nearly as large as marine deposits,¹ are only rarely preserved in a permanent column of sedimentary rocks; they represent rather the temporary incidents of transportation. Of the sediments deposited more or less permanently, a large part, possibly as much as 90 per cent, is deposited in the sea,² and the rest mostly in continental areas, including flood plains,

¹ BARRELL, JOSEPH, Marine and terrestrial conglomerates, *Geol. Soc. America Bull.*, vol. 36, p. 284, 1925.

² BARRELL, JOSEPH, Relative geological importance of continental, littoral and marine sedimentation, *Jour. Geology*, vol. 14, pp. 355-356, 1906. The "littoral" zone is restricted by Barrell (*op. cit.*, page 423) to the zone between high and low tide and receives little permanent sediment. Some have used the term to signify the shallow marine zones.

deltas and lakes. The structural and paleontologic criteria of continental deposition have been considerably studied but are complex, because continental deposits include a variety of agents and conditions of deposition; glacial, eolian and aqueous sediments are often reworked by first one agent and then another until the result is an uncertain mixture. Marine deposits are, however, quite free from mud cracks (Fig. 205),

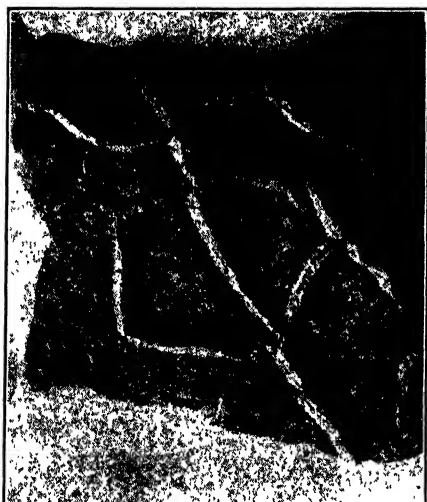


FIG. 205.—A cast of mud cracks in the Medina sandstone, New York. About one-sixth natural size.

rain prints, footprints, and other signs of exposure to air during deposition. An attempt is made to tabulate some of the chief criteria, in spite of their rather unsatisfactory nature. Further readings are also listed.

Both marine and continental areas of deposition may be subdivided according to restricted environments. Marine divisions include (a) abyssal, (b) bathyal, and (c) shallow. Continental divisions include (d) alluvial and colluvial, (e) lacustrine, (f) desert and eolian, (g) glacial, and (h) residual.

a. In the deep sea the dominant sediments are red muds and oozes, but it is doubtful whether many exposed rocks originated at such depths. Vertebrate fossils of extinct species are dredged up from the deep seas, showing that the rate of sedimentation must be extremely slow, probably only a few feet since the Tertiary.¹

b. At moderate depths, 500 to 600 fathoms, where the continental shelf breaks off into the deep sea, the dominant deposit is "blue mud," said to have iron sulphide in it. As the water becomes shallower the chances of persistence of some CaCO_3 become better, but most limestones are accumulated in shallow seas (page 326). Few ripple-marked sediments occur at such depths, but the beds are extensive and uniformly fine grained.

c. Shallow waters in the ocean include those from the shore to a depth of about 100 fathoms or to the edge of the continental shelf. Here are deposited the main bulk of sands and much of the clays and other mechanical sediments. Some of the fragments, those of calcareous organ-

¹ Smithsonian Inst. Rept. for 1899, Appendix by Sir John Murray, pp. 316-318, 1901. But Twenhofel, in Geol. Soc. America Bull., vol. 40, pp. 385-402, 1929, suggests larger quantities of sediment.

isms, are derived from the sea, but far the larger portion is derived from the land. The organisms are commonly recognizable as marine forms, though fragments of land forms may be washed in. Foraminifera are usually regarded as marine.

Special modifications of shallow-water deposits are distinguishable as estuarine, lagoon, Mediterranean, and main sea areas, but the criteria for such formations are probably more structural than petrographic.

Where a large part of the material supplied by the land is in solution, so that the water is clear, and especially where such clear waters are warm and in enclosed lagoons, limestones form the chief sediments of the shallow sea.

The general gradation of sediments from the shore to deeper water is toward finer grain. Beyond 100 fathoms most sediments are oozes and muds. Greensand (glauconite) is characteristic of the "mud line" but may precipitate in sands and limy deposits. It is found mostly where the land supplies only a little mechanical sediment and where the temperature of the waters is variable.

It should be noted that sediments of shallow seas commonly make up formations of greater thickness than the 100 fathoms mentioned as the depth of such seas. It is clear that the sea would have been filled with sediment, if there had not been a progressive change of the relative position of the rocks below the area of deposition. In places where the sediment is found to be many thousands of feet thick, the shallow seas must have been maintained by a sinking. Thus is developed a geosyncline, and in this the younger beds tend to be more simple and uniform in materials than the lower, older beds, probably because of the lower relief and more stable conditions on the land area after long erosion. Great uniformity in petrography is to be expected at a distance from shore in marine, well-sorted sediments that were formed late in a period of deposition. Sediments among islands and in epicontinental seas are more varied, but the most extremely varied sediments are probably those deposited in continental areas and supplied by a variety of older rocks.

A certain amount of sorting and concentration of minerals is characteristic of deposition or the last stages of transportation. The heavy residual minerals seem to be obtained mostly from sands from $\frac{1}{2}$ to $\frac{1}{16}$ mm. in diameter. They may be present in smaller quantities in coarser sediments as inclusions not easily separable, and in finer sediments as dust. Streaks of sand containing a natural concentration of such heavy residuals, although they may not have much variety, are taken as signs of wind and water currents, fluvial or shore conditions of deposition. Marine sands are said to have more variety in heavy residual minerals than terrestrial sands.

Unconformities show petrographic as they do paleontologic breaks.

Deltas are defined as formations built up at the edge of a permanent body of water. They are transitional from marine to terrestrial deposits. Waves and tides tend to destroy the deltas, but some grow in spite of that opposition. Small deltas in lakes may have steep foreset beds (Fig. 206), but the larger ones slope more nearly 1 in 150. Such slopes are rarely divisible into "topset," "foreset," and "bottomset" beds. Text

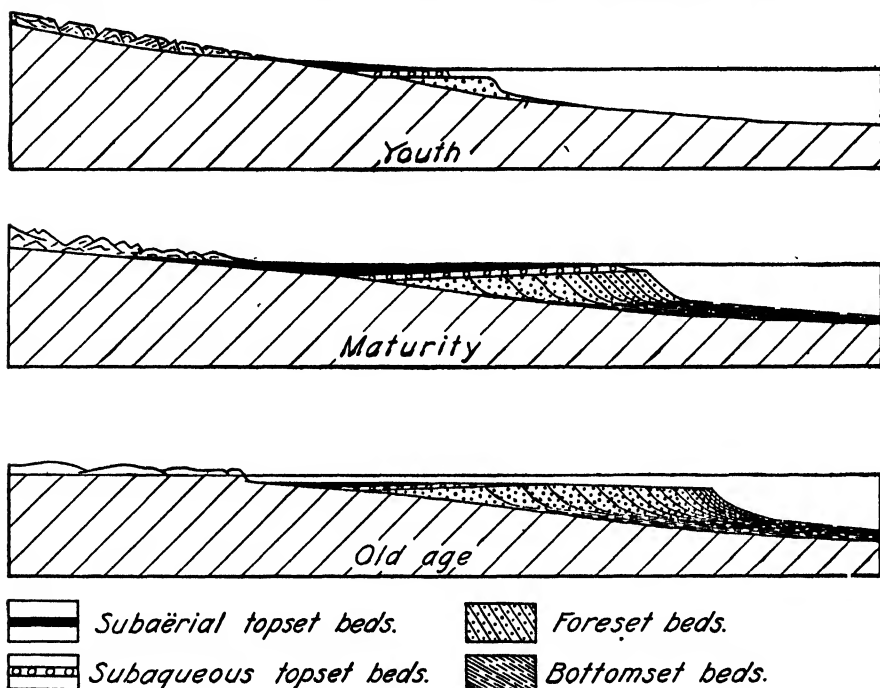


FIG. 206.—Stages in a delta cycle in a large river during a period of stationary crust. (After Barrell, *Geol. Soc. America Bull.*, vol. 23, p. 396, 1912.) Vertical scale greatly exaggerated.

book figures of delta structures use an exaggerated vertical scale. Variability is the chief result of the deposition of deltas near the sea level. Vertically, laterally, in minerals, textures, color, and fossils delta beds alternate rapidly. The organic matter carried by the rivers may be abundant in one bed, whereas in the next, long exposure to air has entirely oxidized it. Land fossils prevail in the topset beds, and occur in the bottomset beds along with marine forms. Nearly all delta beds are notably fossiliferous.

d. Rivers above their deltas deposit chiefly flood plains and alluvial fans. The larger fans and piedmont slopes may be very flat and hardly distinguishable from the plains. Beds of river gravel may become thicker than is expected from gravel beds of marine origin. Playas are

basins receiving deposits from streams that do not carry sediments out of the basins. River deposits are poorly bedded and have cross-bedding characteristic of currents and channels. The edgewise conglomerates of streams have platy pebbles dipping up stream, in contrast to those of beaches which have platy pebbles normal to the main bedding. Fossils are not abundant and if present are recognizable land forms or fresh-water forms. Heavy residual minerals are of local origin and show little variety.

Glacial streams deposit eskers and kames in the form of ridges, also outwash, which more resembles an alluvial fan, but which may have an occasional kettle hole, where an ice block melted after the sediment was deposited.

e. Lakes receive sediments, which are deposited under conditions somewhat different from those controlling marine sedimentation. No concentration of sea salt is present to coagulate the fine clay and colloids. The settled clays may show a lamination, especially in glacial lakes, from which estimates of geologic time may be made. "Varved" clays show an annual gradation from sandy to claylike material, and are believed to form in cold water that is not salty (Fig. 167).

f. Deserts and eolian sediments are characteristically formed where there is a lack of vegetation to bind the soil. This may be due to aridity, to cold climates after glaciation, or to burial under fresh sediments on flood plains and tidal flats. Eolian sands are of small extent, but the finer loess may extend hundreds of miles. Wind-blown sediments are not so well stratified as water-borne sands, but there is better sorting by size of grain deposited. The ripple marks made by wind have a wave length greater than water-wave ripples of the same amplitude. Eolian sands formed in rigorous climates may show rounded grains of easily decomposed minerals, and more-abundant rounded feldspars than are expected in water-borne sands. Dreikanter are characteristic if found, and associated precipitates include salt and gypsum. Even the small grains, below .03 mm., become rounded and frosted by wind action.

g. Glacial *débris* is deposited by melting ice. When glacial flow is in equilibrium with melting, the sediment is deposited in great terminal moraines, but the more widely spread ground moraine is deposited during the retreat of a continental sheet. These deposits may, of course, be modified by the action of lakes and streams, related in various ways to the melting ice, and by wind action. The *débris* is poorly, if at all, sorted or bedded and has uneven thickness and few fossils (pages 314 to 316).

h. Residual sediments accumulate chiefly where (1) rock decay is rapid and (2) mechanical erosion slow. Decay is favored by the solubility of minerals present and by the porosity of the rock. Moist regions of moderate to low relief with some vegetation seem favorable. The chief agent of decay is water for both hydration and solution, but CO_2 , O,

H₂SO₄, and possibly soil acids may be more favorable in some places than others. Some mechanical disintegration without erosion is favorable to decay.

Residual deposits are poorly if at all bedded but form lenses and pockets grading down to other rocks. Quartz grains commonly remain with the form they had in the original rock—some euhedral, some interfered with by their neighbors but few rounded. Micas may be hydrated or leached but commonly retain the form of "books." Heavy residual minerals show little variety, having been derived from a single local source.

CRITERIA OF PLACES OF DEPOSITION OF MECHANICALLY TRANSPORTED SEDIMENTS

Field structures.

Unconformities, yardangs, and overlap commonly indicate the beginning of a marine deposit.

The normal sequence of a geosyncline indicates marine shallow deposition.

Deltas indicate the mouth of a river and both continental and marine shallow deposits.

Fans are continental.

Channels, lenses and pockets, and abrupt changes in thickness are more likely in continental deposits. Uniform beds that thin out gradually or grade into others are more likely marine.

Conglomerates with imbricate pebbles are terrestrial, especially if extensive and of constant inclination.

Thick conglomerates are continental, many are fluvial.

Varved clays formed in fresh, cold water.

Association with sediments of known place of deposit, such as salt and gypsum in arid lands and coal in moist cool land, is suggestive.

Structures.

Mud cracks are mostly continental; crusts may curve differently in salt and fresh water.

Mud balls and rolls, footprints, rill marks and swash marks are continental, largely delta features.

Cross-bedding by its uniformity (or lack of it) and the curvature or flatness of base may indicate whether the sand was deposited by waves or currents of water or by wind (see page 316).

Ripple marks by their size indicate depth of water in wave action. Current and wind ripples less symmetrical. The eolian ripples have large wave length in proportion to amplitude.

Textures.

Size of grain is less than 5 mm. in large deposits of wind-blown material. The range in grain sizes varies for aqueous sediments but is complex. The minerals in each grain size may also be related to place of deposit. Wind-blown material is best assorted and restricted to one size; marine deposits are next best; pluvial and fluvial deposits are poorly assorted and glacial deposits unassorted.

Shapes and Character.—Euhedral and anhedral grains, and "books" of altered mica are mostly residual; irregularly angular to subangular may be pluvial or glacial; well-rounded grains, especially if less than 1 mm., suggest eolian deposits; dreikanter and einkanter, frosted or pitted sands and desert varnish or case hardening suggest eolian deposits; striated and faceted pebbles are glacial; all mostly continental. Fulgurites suggest continental exposure. The more abundant marine sediments have subangular to rounded grains but may rework material from the land with little modification. Beach pebbles either lacustrine or marine tend to be discoid.

Fossils.—Forms characteristic of land, fresh water and salt water.

Fossil foraminifera are said to be marine. Whole and unbroken forms are probably in place of deposition, as are roots in place; broken and especially rounded fragments are less clearly signs of location.

Materials.

Minerals in detail.

Partly decomposed rock fragments are residual, continental.

Red clay with ash and siliceous fossils indicates deep-sea deposition.

Glauconite forms in the sea near the gradation from sand to mud.

Associated salt and gypsum suggest continental deposits.

Coal may form in bodies of salt water or fresh water.

Pure quartz sands are mostly marine (reworked).

Black shale, oil rock, etc. are mostly marine at moderate to shallow depths.

Heavy residual minerals concentrated in streaks suggest deposits that are continental—fluviatile or at the shore of the sea. The greatest variety of residual minerals is to be expected in seas, near several bodies of land; the least variety in continental deposits.

Variegated colors in sediments suggest continental deposits.

The oxidation of iron suggests continental deposits.

Any lack of decomposition in the clays (or clay matrix) suggests glacial continental deposits.

The depth of marine deposition may be estimated by the proportion of carbonate, ranging from more than 90 per cent in some shallow-water deposits to perhaps 50 per cent at 2,000 fathoms and none below 3,500 fathoms.

Uniformity of sediments over wide areas suggests marine deposits at considerable depth; the least uniform beds are probably continental.

Summary of Mineralogic Effects in Mechanical Deposition.—The several places of deposition of mechanical sediments have only general mineralogic effects. The muds are characteristic of deeper sea deposits than sandstones (and organic limestones). Glauconite is abundant at the "mud line." A few other mineralogic characteristics of certain shallow-water deposits have been suggested, but more work is needed for definite conclusions. Glacial sediments especially clays are said to be less weathered than other clays. The altered volcanic ash bentonite is mineralogically distinguishable.

Diagenetic effects in mechanical sediments are described under chemical deposits.

Structures of Mechanical Deposition.—An outline of sedimentary petrology is not a proper place for detailed study of structural geology, but the structures made by mechanical deposition constitute a fascinating field for research, some of which is well under way.

Bedding, as the characteristic of sediments, is commonly attributed to changes in texture and changes in material. Such changes result from a change in source, or a change in the current carrying the sediment or a change in rate of settling (as for example by salting out a suspension). Some bedding is modified if not entirely made over by waves and tides reworking the deposited material. It is characteristic of seasonal variations of texture that the coarser sediments of the flood season merge gradually upward into finer sediments in the alternating season of low water, and this is followed by an abrupt change at the beginning of the next flood season. The repetition of such seasonal gradations and changes makes the "varved" clays of glacial and other lakes (Fig. 167).

Lamination, usually a color banding and tendency to split, is the more minute variation not involving any great textural change. It may be shown by experiment that even when a clay is supplied at a uniform

rate and in uniform sizes, some lamination develops by its deposition and settling. The tendency of clays to split is "fissility."

Cross-bedding results from deposition on locally sloping surfaces (Figs. 174 and 175). These slopes may be a result of erosion of channels or deposition on delta slopes or on dunes. These various causes are in some cases distinguishable in the deposited rocks by the straightness or curvature of the base, by the uniformity of slope (or lack of it) and by the repetition of the structure in later beds.

Ripple marks (Figs. 176 and 207) are irregularities in bedding developed by waves and currents—some of which cross and combine with one another in complex fashion. Although most of them seem to develop in shallow water or in sand dunes, the best chance for them to be preserved



FIG. 207.—Ripple-marked graywacke sandstone, Pigeon River, Minnesota.

seems to be in the flood plains of rivers. The origin of some ripples can be determined by a detailed study of size, form, symmetry, etc. They may even suggest the depth of water or the direction of the nearest land mass.

Wind and water-current ripples are asymmetrical, but those formed by wave action are symmetrical. It is said that the ratio of wave length to height distinguishes wind and water-current ripples. Ripples under water have wave lengths three to six times as great as the vertical distance from crest to trough; those formed by wind, twenty to thirty times.¹

Mud cracks or sun cracks (Fig. 205) form in playas and flood plains, less commonly on the shore zones of lakes and seas. It is said that the warped layers curl up in fresh water and curl down in salt water, but the feature must be used with caution.

¹ BUCHER, W. H., On ripples and related sedimentary surface forms, *Am. Jour. Sci.*, 4th ser., vol. 47, pp. 147-216; 241-269, 1919. Few ripples on dune sand have wave lengths greater than 4 inches.

Pit and mound structures and crater structures may result from several different processes—rapid fall of a coarse grain into mud, a raindrop, a hailstone, or the rise and bursting of a bubble.

Miscellaneous structures include rill marks, swash marks, footprints, trails, beach cusps, clay galls, clay boulders and folds from subaqueous gliding (page 341). Later structures include joints, faults, folds, etc. Clastic dikes are said to be derived generally from lower formations and to be pressed up by the weight of the rock load or by gas pressure. The sand grains of dikes are well rounded and were no doubt lubricated by some fluid while being injected. Larger structures, involving erosion as well as deposition, are the unconformity, disconformity and overlap.

Several of these structures may persist during deformation and tilting and determine the original position of deformed and overturned beds.

Criteria are given by Tanton for the tops of beds¹ after some deformation and metamorphism, and by Leith for unconformities in a series of beds.²

ORGANIC DEPOSITION

The organic deposits of sediments were grouped mineralogically as (1) calcareous, with more or less magnesia, (2) siliceous, (3) phosphatic, (4) ferruginous, and (5) carbonaceous. Of these the calcareous carbonate rocks are of much the greatest importance in the shoal waters of the ocean.

The processes of organic deposition are distinguishable into three groups—those in which the functioning organic material is accumulated, those in which certain hard parts (skeleton or shell) are accumulated, and those in which the functioning organism causes a reaction that precipitates some substance from the solution around it. The study is complicated by the fact that most substances formed by organic action may also result from reaction or evaporation.

Organic Deposition of Calcium Carbonate.—The calcium for sediments is carried in solution as sulphate or as bicarbonate, deriving its CO_2 from the air. The organisms commonly recognized as forming limestones are noted on page 290. The chief organic deposits of calcium carbonate have been supposed to be shells and internal structures of sea animals, but it is now believed that there is as much precipitation by bacteria, algae and other microscopic organisms as by the larger forms of life. It should be noted that some of the precipitates result from decomposition of calcium bicarbonate when plants remove CO_2 from carbonated water.

Places of Accumulation of CaCO_3 .—1. Coral reefs grow in Recent times in warm (68°+F.), shallow water, normally at depths of less than 25

¹ TANTON T. L., Determination of age relations in folded rocks, *Geol. Mag.*, vol. 67, pp. 73–76, 1930.

² LEITH, C. K., "Structural Geology," rev. ed., pp. 278–282, 1923.

fathoms but possibly down to 45 fathoms. The reefs may have only a small percentage of coral.¹ It is said that corals are fourth in order of abundance among the organisms in the reef at Funafuti.

2. Reefs of other organisms occur at similar depths.

3. Bedded fossil deposits—crinoids, foraminifera, ostracods—form coquina and oozes. Globigerina oozes cover about 50,000,000 square miles of sea bottom. Foraminifera in fossil forms are said to be wholly marine.

4. Shell colonies, forming local masses in shales.

5. Fresh-water lakes have algae, snails, etc., forming marls.

Balance of Solution and Deposition of CaCO₃.—Organisms secrete CaCO₃ from water even when the water is not saturated, and when the organism dies, the CaCO₃ may begin to dissolve. This solution, according to conditions prevailing, may be so slow that other deposits cover and protect the early ones or so rapid that no limy material accumulates.

The rate of growth of a limy deposit varies largely with the supply of food, the supply or concentration of calcium and the freedom from muddiness and unfavorable forms of life.

The rate of solution, on the other hand, seems to be largely related to depth in the sea, though at the surface the water is more nearly saturated in some places than others. The depth relation is apparently due to the fact that CaCO₃ dissolves mostly as the bicarbonate, and that pressures deep in the sea hold large quantities of CO₂ in solution (page 333). The low temperature also favors such solution, and the supply of CO₂ seems to be increased by the decay of organisms that have settled into the depths. Whatever the reason, CaCO₃ is relatively scarce in the deeper deposits.

Ooze at 250 fathoms has 86 per cent CaCO₃.
Ooze at 2,200 fathoms has 46 per cent CaCO₃.
Ooze below 3,500 fathoms has no CaCO₃.

Again, in coral reefs (as in shallow sediments) the evidence of a sinking of the bed of the ocean seems to be conclusive.

Composition of Deposited Carbonates.—The composition of abundant organically deposited carbonates is indicated by the following table. Carbonate rocks richer in magnesium carbonate are probably the result of diagenesis or later metamorphism, though it is not impossible that in past periods some organisms may have secreted somewhat different carbonates from those secreted by organisms now living. Warm waters and therefore regions near the equator are said to favor magnesian shells.²

¹ SOLLAS, W. J., Funafuti; the story of a coral atoll, Smithsonian Inst. Ann. Rept. for 1898, pp. 404–406, 1899.

² CLARKE, F. W., and W. C. WHEELER, The inorganic constituents of echinoderms, U. S. Geol. Survey Prof. Paper 90L.

TABLE XVIII.—MAGNESIUM CARBONATE CONTENT OF LIMY SECRETIONS OF VARIOUS ORGANISMS
(After Clarke)

	Per Cent MgCO_3
Foraminifera	3.67 to 11.22
Sponges	5.37 to 8.00
Aleyonarians	6.03 to 15.73
Echinoderms	5.99 to 14.08
Brachiopods	0.49 to 8.63
Mollusks	0.00 to 2.58

Textures.—The textures of organically deposited CaCO_3 have been outlined in the descriptive part of the text. Primary textures are organic but there is a special tendency for carbonates to recrystallize and become cemented, and it is also common for such material to be reworked by waves and wind into mechanical calcareous sediments both arenites and lutites.

Organic Deposition of Silica.—Both fresh and salt waters have enough dissolved silica to supply considerable beds of siliceous sediments.

SiO_2 = 12 per cent of river solids, perhaps 20 parts per million.

SiO_2 = 0.01 per cent of sea salts, 2 to 4 parts per million.

Silica is taken into solution by $\text{MgH}_2(\text{CO}_3)_2$ and other waters and may be protected from coagulation by organic colloids.

The organisms that deposit silica are diatoms, radiolaria and sponges. The growth of these organisms is conditioned largely by the temperature of the sea. As the diatoms in particular float in the upper parts of the sea, their remains may fall into the deep sea as well as the shallow sea. This makes it likely that considerable masses may be free from shore sediments, though others may be mixed with shale. Those masses now exposed on land were probably formed at shallow depths. The enormous cherty formations constituting iron protores near Lake Superior have locally some remains of algae and bacteria, which may have had a part in their precipitation (Fig. 208).

Organic Deposition of Iron and Manganese Oxides.—Weathering does not commonly leach iron, but in some swamp waters iron may be leached from silica. Laterites are red from iron minerals, and many iron ores are residual enrichments. Iron in a ferrous state is transported as sulphate and bicarbonate, but in surface waters these salts rapidly become oxidized, and iron tends to precipitate unless organic colloids serve to "protect" it.

Iron is carried by some spring waters, by river waters usually in small amounts and by the sea water also in small amounts. At intervals in geologic history great iron formations have been deposited, suggesting a supply of more concentrated solutions, possibly igneous, possibly organic.

Bacteria are active in precipitation of iron¹ (Fig. 208). Protozoa, algae and fungi are also active but less important. A similar set of organisms precipitate manganese.² It is probable that some iron deposits may result from the "coprecipitation" of iron with some other colloidal matter.³ Mineralogically most of the iron is in hydrated ferric oxides, but some sulphides result from organic action.⁴

Large iron deposits probably formed in shallow seas, but bog ores form in swamps and lakes.

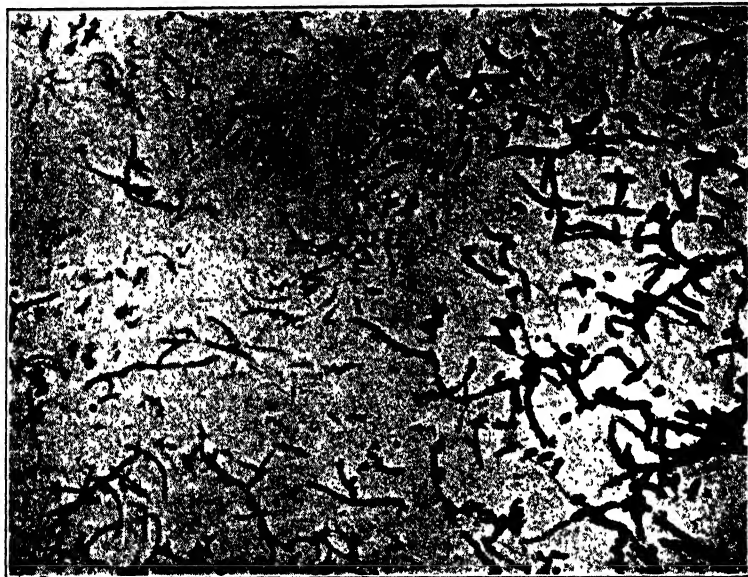


FIG. 208.—Photomicrograph of forms that appear to be bacteria and algae in a chert pebble, probably a pebble from an Archean rock. (After Gruner.) The dark tubes are ferruginous. $\times 75$.

As to texture, the precipitates are microscopically fine grains forming sheaths or fibers, etc. It has been suggested that some granule textures with about $\frac{1}{16}$ -inch granules and some oölitic forms may be due to organic deposition. The greenalite and chamosite and related rocks are wavy bedded and laminated, grading into intraformational conglomerates as well as into other sediments.

Organic Deposition of Phosphate.—Most phosphatic remains are probably the bones of vertebrates or the shells of crustaceans and brachio-

¹ HARDER, E. C., Iron depositing bacteria and their geologic relations, U. S. Geol. Survey Prof. Paper 113. J. W. GRUNER, Algae believed to be Archean, Jour. Geology, vol. 31, pp. 146-148, 1923.

² THIEL, G. A., Precipitation of manganese from meteoric solution, Am. Jour. Sci., vol. 7, pp. 457-472, 1924.

³ BOSWELL, P. G. H., Geol. Mag., vol. 67, p. 371, 1930.

⁴ THIEL, G. A., Biochemical reduction of sulphate waters, Econ. Geology, vol. 25, pp. 242-250, 1930.

Pods. The coprolites and the guano deposits from birds on oceanic islands are phosphatic—there are phosphatic bones in the birds' diet. The percentages of phosphate in remains of other organisms are as follows:

TABLE XIX.—PERCENTAGE OF $\text{Ca}_3(\text{PO}_4)_2$ IN REMAINS OF VARIOUS ORGANISMS
(After Clarke)

Lingulas	91.74
Discinisca	75.17
Alcyonarian	8.57
Crabs about	15.00
Shrimps about	30.00

Most of these organisms deposit phosphate in the shallow seas or along shores, but bones and teeth may settle in the deep sea. They probably derive the P_2O_5 ultimately from the sea water, which is said to have about 0.015 per cent of P_2O_5 in solution.

The phosphate in nearly all of these deposits is collophane, and as deposited in Recent times this is mixed with carbonate to such an extent as to indicate that most good phosphate rock is altered and enriched after deposition. This belief is confirmed by the fact that nearly all phosphate deposits are well cemented. See Diagenesis and the list of further readings.

Deposition of Organic Compounds.—Plants seem to have been preserved as actual carbonaceous matter much more abundantly than animals. Possibly animal matter contributed to petroleum, but plants are now regarded as important sources of this product also. Coals and oils of the several kinds are the chief carbonaceous deposits.

Plants derive their carbonaceous material largely from the air and partly from ground-water solutions of carbon compounds. The composition and texture of the accumulations of vegetation vary with the nature of the organism, but the variety of forms is too great to be considered here in detail. Many beds of coal have structures indicating tree trunks of considerable size, but some surprisingly thick deposits of cannel coal seem to contain only microscopic spores. See list of further readings.

The nature of the vegetation largely influences the processes that modify the accumulation later, but the conditions of transportation and deposition may be even more important. It is only in exceptional cases that vegetation accumulates on land surfaces, but a growth that falls into water or is transported to a body of water may be protected from decay.

Possibly some bodies of coal have accumulated where they lie as prolonged forest growths in wet regions. An occasional forest fire, caused by lightning, seems to have interrupted the growth and left a thin bed of charcoal—"mother of coal"—though certain beds of charred appearance may have resulted from dry rot. It is possible also that

swamps quite free from forest growths may support a growth of small plants great enough to make important deposits of peat.

There is evidence that many other coal beds have resulted from transportation of vegetable débris and its accumulation in lakes or ocean basins. The microscopic forms may be partly wind borne, but as water is an important element in their preservation and much of the coal is in large fragments, it seems probable that streams were the most important agents of transportation. Some very thin deposits of coal have been traced several hundred miles.

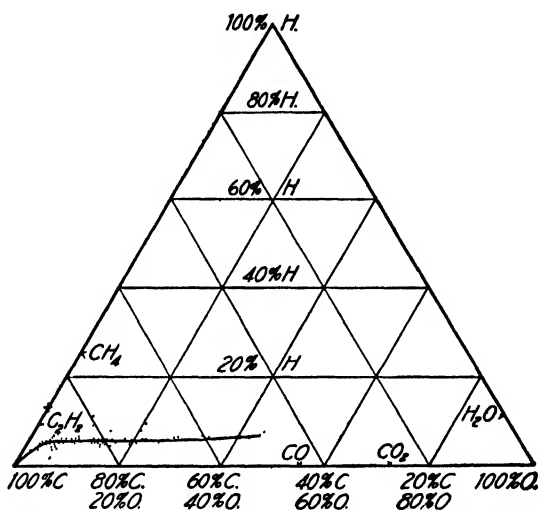


FIG. 209.—Diagram of the coal series. None of the coals (ash and water free) are rich in hydrogen, and they range in oxygen from 50 per cent down, and in carbon from about 50 per cent up to 100 per cent. Spots represent individual analyses and the curved line shows the trend of alteration from wood at the right end of the line to graphite at the left.

Vegetation has not been known to evolve into coal except after burial under other sediments. The process of alteration of swamp deposits to peat may be the initial step of coal formation, but it is a relatively minor step. The softer organic material by long standing seems to develop a jelly or matrix, in which larger, more resistant fragments are suspended, and the combination accounts for some textural features seen in average coal. After burial the progress of alteration is indicated by the series from lignite through the higher grades of coal to graphite (Fig. 209).

The change of vegetable tissue to coal involves volatilization and a variety of chemical reactions. The result varies with (1) the nature of the original vegetation, (2) the partial decay of the material before burial, (3) the time since burial, (4) the depth of burial, (5) deforming pressures, (6) cracks that allow the escape of volatile matter, and (7) heat. Experimental work indicates that shearing generates and drives out some volatile hydrocarbons. The more-volatile materials eliminated from

coal seem to be such as might produce oil and gas. Many coal mines have difficulty with an abundance of gas closely resembling that in the larger reservoirs of natural gas. Petroleum accumulations, however, can rarely be traced to any direct genetic relation with coal. Much more organic matter is disseminated in sediments than is concentrated in formations like coal.

Petroleum may be formed by a variety of inorganic reactions, but the commercial accumulations seem to have been formed from organic remains and occur wholly in sediments. In some deposits plant remains

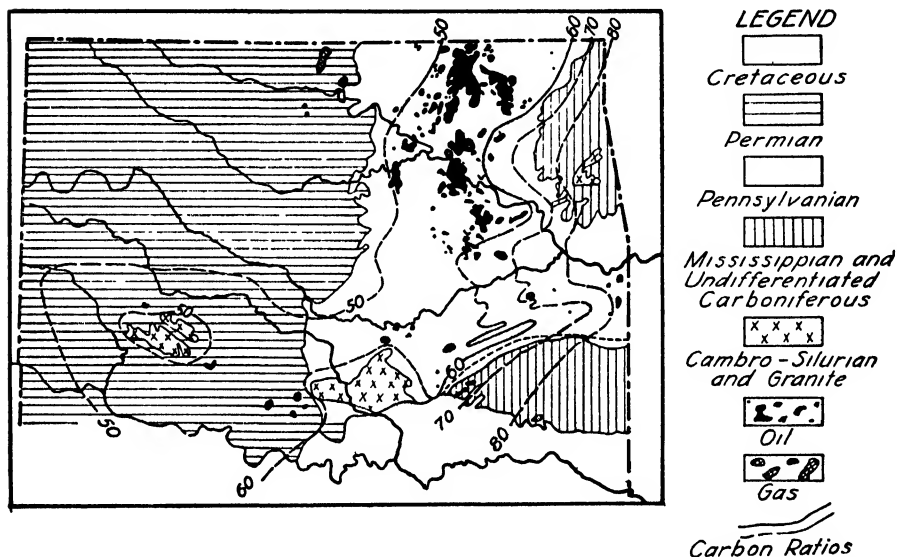


FIG. 210.—Map of Oklahoma, showing oil and gas pools and lines of equal volatile matter in coal. (After Fuller, *Econ. Geology*, vol. 15, p. 232, 1920.)

may predominate, in others mollusks, fishes and microscopic animals may contribute more than plants. The varied nature of petroleum in different localities no doubt reflects differences in the original organisms, as well as in the later history.

The process of evolution of gas and liquid from organic remains is complex and is much modified by conditions of heat and pressure. It is important to note that the progress of metamorphism in the coal and shale beds of a district is a valuable guide to the probability that gas and oil are retained in the structures of accompanying sediments (Fig. 210).¹ No great oil fields are to be expected in association with the more highly altered coals (fixed carbon more than 65 per cent of the pure coal) or with the harder shales. The change from lignite to anthracite is roughly parallel to the change from clay to slate.

¹ WHITE DAVID, Washington Acad. Sci. Jour., vol. 5, p. 212, 1915. E. R. LILLEY, Coal as an aid to oil exploration, Eng. Min. Jour., vol. 117, pp. 1010-1012, 1924. J. H. WILSON, Lithologic character of shale as an index to metamorphism, Am. Assoc. Petroleum Geologists, vol. 10, pp. 625-633, 1926.

The oxidation of petroleum yields solid and semisolid asphaltic deposits, which are more easily considered as rocks than the liquid and gaseous hydrocarbons. The oxidized residues may fill fractures, cement breccias and sands, or accumulate at the surface of the ground, forming asphalt lakes.

CHEMICAL DEPOSITION

Aqueous rocks may form by precipitation from a variety of solutions—igneous emanations, ground waters of meteoric origin, lakes and ocean waters. The ocean, the largest body of these solutions, receives its mineral content largely from river waters, which dissolved their mineral matter partly underground and partly during weathering near the surface.

Evaporation.—On evaporation of sea water a series of minerals precipitate. When the volume is reduced to about 50 per cent of the original,

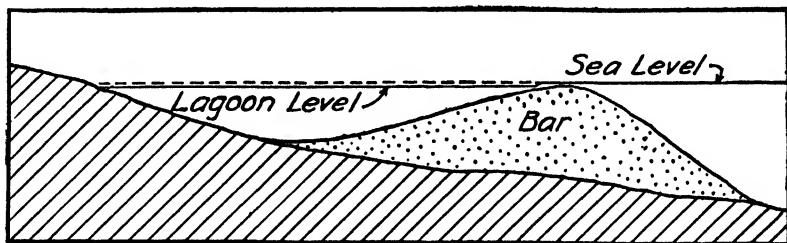


FIG. 211.—Diagram of the conditions under which salt and gypsum may form thick beds. Ochsenius's bar theory.

the only precipitates are (1) Fe_2O_3 and (2) CaCO_3 . When four-fifths of the water is removed, some (3) gypsum precipitates; (4) salt, NaCl , precipitates only after the volume is reduced to about one-tenth of the original. These several precipitates overlap a little, but some beds may be formed of fairly pure minerals, with only a small gradation zone to the next mineral. After the beginning of salt precipitation (5) anhydrite forms rather than gypsum. Later there are precipitated (6) magnesium sulphate, (7) the double sulphate of magnesium and potassium, (8) magnesium bromide, (9) the double chloride of magnesium and potassium, (10) a hydrous magnesium chloride.¹

The sea contains about 3.6 per cent of salts, of which 78 per cent is NaCl . Thus, if 1,000 feet of sea water is evaporated, the *salt* layer is 15 feet thick and *gypsum* less than 2 feet thick. To produce the many thick beds known in the geologic column we can not imagine a simple drying up of some very great bodies of sea water.

Very important salt deposits seem to have formed from sea water in special localities such as in a lagoon or behind a bar, with only a small

¹ CLARKE, F. W., Data of geochemistry, U. S. Geol. Survey Bull. 770, p. 220, 1924; data from Usiglio.

channel connecting with the sea (Fig. 211). In such an area concentration by evaporation is reinforced by constant additions of salt water from the sea. Lakes with no overflow and desert basins may also receive large supplies of salt.

Thick beds of gypsum may also be explained as deposits in lagoons, or by replacement, or by the drifting of gypsum sediment into low places by waves or into sand dunes by wind.

Few chemical deposits occur in the deep sea, although some nodules of manganese oxide, of phosphate or of glauconite may be attributed to chemical action. Those in the shallow sea are more important and include aragonite needles and possibly some oölites and ancient dolomites. The solubility of CaCO_3 in sea water at 17°C . is about 191 parts per million, and the sea contains about 121 parts per million. A moderate evaporation would therefore cause precipitation. The solubility may be largely modified by CO_2 . In pure water CaCO_3 dissolves only very slightly—16 parts per million, but with CO_2 present at one atmosphere pressure, the amount is increased seventy times by the formation of bicarbonates.

The precipitation of CaCO_3 from sea water has been the subject of several investigations. As noted above it may occur as a result of evaporation, but few limy muds seem to have this origin. Where CaCO_3 seems to be depositing inorganically, the water probably contains more lime than the average sea water, and as the material is largely held in solution by an excess of CO_2 its precipitation is caused by a loss of CO_2 . This loss may be the result of agitation, barometric changes, rise in temperature or the use of CO_2 by other compounds or by organisms. See further readings on these reactions.

The evaporation of waters other than sea water has been little studied. Evaporation in lakes and at the surface of the soil in arid regions forms deposits of NaCl , sulphates, potash salts, borates, and others. "Alkali" may become so concentrated in soils of large areas as to ruin their fertility. This alkali is probably sodium hydroxide rather than the carbonate. It is said, however, that terrestrial salts commonly have sodium carbonate, which distinguishes them from marine salts. Other deposits from solution in semiarid regions, where ground-water circulation and capillary action bring solutions to the surface, include bauxite, iron crusts, caliche and the silica of the "case-hardened" sandstones.

The textures of evaporates are determined largely by the nature of the minerals in them, though some of the rocks may be reworked into diagenetic modifications. Crystals are normally intergrown and have very little porosity. It is probable that precipitation that occurs where waves agitate the crystals may produce oölitic structures. Wind and wave action on some precipitates like gypsum may develop fragmental textures. The textures of caliche and borates range from that of soil with a little cement to masses of salts with only traces of the sand left in it.

Reaction of Solutions.—Few bedded sediments are to be attributed to reactions in solutions. Possibly some cherts result from a coagulation of colloidal silica in salt water and possibly some bedded iron minerals may

be chemical precipitates. The action of oxygen from air on an iron solution probably occurs as the oxygen dissolves and solutions react.



The textures of such bedded precipitates are normally very fine or colloidal, unless reworked as many are by diagenesis. The criteria of a former colloidal state are listed, but many doubtful cases are found (Fig. 212).

More definitely, we believe that ground waters with unlike minerals dissolved in them may react to precipitate vein fillings and cements in



FIG. 212.—Structures in a rock high in silica, suggesting colloidal deposition and later reworking of the material into cryptocrystalline structures. Chalcedony from Quartzburg, Oregon. $\times 25$.

sediments. This action is especially likely if magmatic and meteoric waters mix. Hydrolysis may be a factor in some precipitations such as that of limonite. The effect of mingling is proved by the fact that ore shoots form especially at the intersection of fractures.

CRITERIA SUGGESTING A COLLOIDAL ORIGIN OF MINERALS

Colloid; definitions.

Material consisting of particles between $1/1,000$ and $1/1,000,000$ mm. in diameter. Suspensoids—solid particles colloiddally dispersed in liquid—do not form gels. Emulsoids—liquid drops colloiddally dispersed in liquids—often form gels. Protective colloids prevent coagulation of other colloidal matter. Peptizers cause dispersion, keeping other material in colloidal solution. Most colloids occur in the zone of katamorphism.

Structures and textures strongly suggesting colloidal origin.

Mammillary, veniform, botryoidal, concretionary, oölitic, pisolitic, nodular, septaria.

Curved banding (Liesegang rings).

Spheroidal forms (except in glass); internal concentric rings where impurities are present, spherulitic structure if substance is pure; concentric contraction cracks; radiating fibers.

Other textures that may result from colloidal origin.

Dendritic forms, three dimensional; stalactitic; sulphides in jasperoid; earthy forms with conchoidal fracture and dull or porcellaneous luster.

Tests indicating colloidal properties.

Plasticity in clays and adsorption of dye or stain.

Isotropism. Anomalous double refraction may indicate a strained gel.

Amorphous conditions of mineral that usually is crystalline. To distinguish amorphous mineral from its crystalline equivalent determine index of refraction (usually lower in amorphous mineral), specific gravity, specific heat, fusibility, solubility.

Absorption of water and regular character of dehydration curve.

Presence of small amounts of many elements not assignable to definite compounds.

Microscopic invisibility of gold in a gold-bearing rock.

Very large crystals may indicate former gel condition of the enclosing mineral?

Metacolloidal minerals themselves are characteristically fine grained.

Minerals occurring as colloids or of colloidal origin.

The hydrous minerals are especially likely to be colloidal.

Sulphur; carbon; silica in the form of opal, chert, siliceous sinter; bauxite and the included oxides of alumina; hydrated iron oxides; pyrolusite; psilomelane; wad; stibiconite; metabrucite; wood; tin; sulphides of iron (marcasite), Cd, Hg, As, Sb, Ag, most of which are found as earthy forms; kaolin and clay minerals; glauconite and greenalite; zeolites, talc; serpentine; meerschauum; chrysocolla; carbonates of Ca, Mn, and Zn; barite; phosphate minerals; niccolite, cobaltite and smaltite; melanterite.

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BOYDELL, H. C., The role of colloidal solutions, etc., Inst. Min. Met. Bull., December, 1924.
HUBBARD, G. D., Colloids in geologic problems, Am. Jour. Sci., 5th ser., vol. 4, pp. 95-110, 1922.

Temperature and Pressure Changes.—Changes of temperature and pressure are the common explanations of precipitation of ore veins and the cements in porous sediments. Onyx may be deposited from cooling saturated silica solutions at springs. The textures of veins, whether formed by reacting or by cooling solutions, range from almost granitoid to amorphous, and many veins have comb structures and vugs.

Snow as an atmospheric sediment may be said to precipitate as a result of change of temperature.

Possibly related to changes in pressure are the effects of agitation of water by waves, which liberates gases from solution, leaving salts less soluble after the gas has escaped.

Recrystallization.—Recrystallization and reactions between solids and solutions, when they occur in sediments, are commonly classed as processes of diagenesis. See the section with that title. There are

replacements, however, that seem to develop a sedimentary or at least an aqueous rock out of entirely different materials. It is said, for example, that a lava below a deposit of guano may be entirely replaced by phosphate rock. Many ore deposits also replace rocks, and nearly any rock may be so replaced.

EQUILIBRIUM OF CONSTITUENTS IN SEDIMENTATION

It is instructive to trace the generalized cycles of changes of state and equilibrium for certain elements and compounds in sedimentation. Space does not permit a detailed study but a few important cases will serve as illustrations.

The most abundant element, oxygen, is not found uncombined in rocks or in meteorites and can not be assumed to have been so in planetesimals. Its concentration at the surface of the earth and in the atmosphere probably results from the emanation of H_2O and CO_2 from magmas, and the action of plant life upon these two compounds. This is the only apparent source of atmospheric oxygen. The losses of oxygen are caused by the oxidation of minerals, and by the oxidation of carbon compounds, part of the latter in animal organisms. It is likely that the atmosphere in early ages had more CO_2 and less O than at present, but approximate equilibrium may have been established long ago.

The ocean contains such vast amounts of gases like CO_2 in approximate equilibrium with the atmosphere, that any conceivable addition to the atmosphere during geologic time would probably have a very slight effect on the proportion of gases in the atmosphere, for the excess gas would soon be largely dissolved in the ocean.

Chlorine is being progressively concentrated from the magmas to the air, to the salt of the ocean. Soda is transferred from the rocks to the streams, to the ocean. Potash, on the contrary, moves from primary feldspar to ground waters, to clays, to sericite and in some slates to a second generation of feldspar.

(See further the cycle of $CaCO_3$, page 326.)

DIAGENESIS AND OTHER CHANGES IN SEDIMENTS

Diagenesis is the alteration of a bed, or a formation consisting of several beds, before consolidation or while in the environment of deposition. It usually occurs before the next bed or formation is definitely deposited over it as a cover. The essential characters of the changes included under this heading are the primary nature of the alterations and their early accomplishment, contrasted with later changes called metamorphism. The two are not always sharply separable. The changes may be mineralogical or structural and include induration, recrystallization, and replacement; leaching, corrosion, oxidation, and enrichment; concretionary action and reworking and structural modifications. Possibly some hydration might be included. (See list of further readings.)

Induration.—Induration usually involves the introduction of some cementing mineral, but in clays the exact nature of the induration is not certain; it may be a sort of pressure cohesion. Pressure brings the grains into close contact and may be the main factor involved. There are in most clays, however, certain colloids, which may serve as binders, and microscopic studies indicate that early in the consolidation of clays there is some recrystallization—sericite forms easily. Diagenetic cementation is commonly effected by ground waters.

Meteoric waters probably act during artesian circulation through long periods of time, but studies in deep mines indicate that their effectiveness is slight except at shallow depths. The main locations of deposition of sands, the bottoms of shallow seas, are probably not very favorable to cementation, for there is little circulation through them. The whole of the liquid in an artesian reservoir moves, though most actively in a direct route and more slowly in the deeper, longer routes. In general the path goes first down, then laterally and finally upward. The upward movement favors deposition. The amount of ground water in the rocks of the crust is much less than was once supposed (page 383).

Magmatic waters are probably more concentrated than meteoric waters and traverse the sediments from the bottom up. Although their volume may not be large, their concentration is great, and the temperature change caused by them is important. Their action is not likely to be early enough to be diagenetic and may be very much later.

The cements are chiefly carbonates, iron oxides and sulphides, and silica in various forms; less commonly there are cements of minerals of hydrothermal types. Clay, hardened by pressure, may also serve as a cement for coarse fragments. Some detailed studies have been made of the stages of hardening of shales.¹ The progress of change appears in specific gravity, porosity, hardness, fissility, crushing strength, slaking, weather resistance, and luster. With an initial porosity of 50 per cent in the first 100 feet, an overburden of 1,000 feet reduces the porosity to about 30 per cent; 2,000 feet to about 23 per cent; 3,000 feet to about 18 per cent; and 8,000 feet to about 8 per cent.

The source of cement may be local or distant. Many of the calcareous cements may be produced by the recrystallization of shells almost as soon as deposited. Similarly, impure clays may recrystallize to sericitic and chloritic cements, and feldspar may alter to argillaceous cements.

The rocks with introduced cements are the fragmental or clastic rocks, the greatest volume of such rocks being of sandy texture. The process is similar in coarser rocks, but fine clays so far interfere with circulation that such cementation is rare. Contact action may cement a clay into a

¹ HEDBERG, H. D., Effect of gravitational compaction on the structure of sedimentary rocks, *Am. Assoc. Petroleum Geologists Bull.*, vol. 10, p. 1058, 1926. J. H. WILSON, Lithologic character of shale as an index of metamorphism, *ibid.*, vol. 10, pp. 625-633. L. F. ATHY, Density, porosity and compaction of sedimentary rocks, *ibid.*, vol. 14, pp. 1-24, 1930.

hornfels of a cherty appearance. "Pipestone" in some quartzite formations seems to be a silicified clay.

The age of the cement may not be so well determined as the time of deposition of fragments. Wetzel suggests that opal cements are Tertiary or later.¹

Recrystallization and Replacement (Metasomatism).—This process is most commonly mentioned in connection with profound changes in structures and minerals that constitute metamorphism. See page 419 where the criteria for recognition of such effects are listed. A milder degree of recrystallization may be a primary process in sedimentation, occurring without so much deformation as is characteristic of differential pressures.

Recrystallization as a process of diagenesis affects mostly the more soluble rocks like limestone. When the growth of a shell bed is not so rapid as to cover the shells at once with other shells, sea water stands in contact with the shells so long that a readjustment occurs within the bed. Small grains disappear, and large grains grow larger in accord with the usual physicochemical rules. By such a process a primary bed of shells may develop crystals about as coarse as those in common marble yet have no signs of deformation in the enclosed fossils. The widely used brown Tennessee "marble" is a striking example.

Metasomatism here includes chiefly dolomitization, silicification, calcification, phosphatization, pyritization, and replacements by FeCO_3 and limonite.

Dolomite is rarely an original deposit (see page 327). A few dolomites seem to be clastic from older dolomite, and others are the results of much later metamorphic or mineralizing activities. Still others, however, in large numbers show clearly that they were changed from calcium carbonate to dolomite before the next formations were deposited.

There is evidence that water may leach CaCO_3 from a slightly magnesian carbonate more rapidly than it leaches MgCO_3 , leaving a residue more porous and more dolomitic. Many dolomites are so porous as to suggest that leaching of this sort has been prominent in their development. Shallow water and abundant CO_2 seem to favor such leaching.

Again, the corals of the Pacific islands are found to be magnesian at depth although scarcely at all magnesian when they grow. It is evident that sea water gradually reacts with CaCO_3 partly replacing it with dolomite.

Borings at other places, however, such as that at Key West, Florida, show no apparent relation between the content of magnesia and depth. Such replacements occur in certain places but not everywhere. The ratio of Mg to Ca in successive geologic periods (page 302) is also significant. See the list of further readings.

¹ Min. Krist. u. Pet. Fortschritte, vol. 8, pp. 101-197, 1923.

TABLE XX.—MAGNESIUM CARBONATE IN BORINGS ON ATOLL OF FUNAFUTI
(From U. S. Geol. Survey Bull. 770, p. 575)

Depth, feet	MgCO ₃ , per cent	Depth, feet	MgCO ₃ , per cent
4	4.23	295	3.6
13	7.62	400	3.1
15	16.40	500	2.7
20	11.99	598	1.06
26	16.0	640	26.33
55	5.85	698	40.04
110	2.11	795	38.92
159	0.79	898	39.99
200	2.70	1,000	40.56
250	4.9	1,114	41.05

Iron ore replacements of the Clinton ore type are diagenetic. The ore has fossils that were certainly CaCO₃ when formed and oölite that probably was the same. Both were largely replaced by iron oxide before the next bed formed, for fragments of the ferruginous rock are found in the overlying limestone in a matrix not impregnated with iron. This replacement was no doubt due to a reaction with ferruginous waters, possibly containing organic salts of iron. Some siderite may also replace calcite.

Silicified limestones, fossils and oölites may also in some cases be formed by early alterations. Silica is derived from spicules, diatoms, etc., and makes the rocks cherty, or almost quartzitic. Loose euhedral crystals of quartz, generated by replacement of limestone, have been known to be concentrated by leaching of the lime, so that the residual quartz grains seemed to make a bed of sand in the limestone.

Guano on some islands causes a replacement of bedrock by phosphate. Several important phosphate rocks elsewhere are also believed to be replacements.

Glaucinite is formed in the sea bottom near the "mud line," probably by reaction of sea water on mixed sediments. Its occurrence in dredgings shows its primary nature, and its common relation to calcareous shells suggests that organic matter may be involved in the process of formation.

The *blackening* of shells, bones and pebbles of limy matter in the bottom of the sea is diagenetic and is partly due to pyrite, partly to manganese oxides.

Pyrite probably develops early as crystals by changes in the black, amorphous iron sulphide that may be present in muds and limestones. Certain fossils in coal and limestone are pyritized. Pyrite crystals in some shales may be a result of organic reducing agents, but the shale is clearly replaced.

Many other types of replacement are known to be subsequent to diagenesis, and the minerals of many orebodies may serve as illustrations.

Leaching, Corrosion, Oxidation, and Enrichment.—Cavities in some limestones seem to represent crystals that have been leached out. A cubic cavity in limestone may have been original salt or later pyrite, but in the absence of an iron stain is most likely to have been salt and to have been leached out very early.

Many limestones show corrosion zones, indicating a cessation of deposition and an actual resorption of the deposited formation. Intraformational conglomerates are the common marks of such an interval. Even sandstones and iron ores may develop such beds. Commonly the pebbles are enriched in phosphate and blackened by iron and manganese minerals, which concentrate during the corrosion. Some phosphate rock is mostly made up of such corroded pebbles.

The alterations that enrich a phosphate bed are of two distinct sorts.¹ A leaching of calcite from phosphatic limestone may leave a surface residual phosphate; carbonated waters seem to have this effect. On the contrary, humic acid waters leach phosphate from such a lean phosphate rock and these solutions may enrich the rock at a lower level.

Stylolites are leached streaks in limestones formed by solution probably under the pressure of an overlying bed. The form of solution surfaces is a series of concavities. Leaching may thus reduce the thickness of beds as much as 40 per cent. See Fig. 186.

The clays below coal beds are in places so pure and free from soluble compounds that there is a strong presumption that water containing organic compounds from the coal plants leached out the impurities of the clay. This probably happened while the organic matter above was growing, but it was so early in the history of the formation that it is essentially diagenetic. The suggestion has been made that some such clays are ash from an oxidized coal.

Ferrous precipitates, such as siderite and greenalite, no doubt promptly oxidize in their superficial layers, if deposition is interrupted in any way. If at the same time fresh water leaches the silica from such a deposit, there may be primary enrichment of an ore bed, though probably such primary leaching seldom forms a thick bed of ore.

Vegetable tissue and peat suffer various primary alterations—partial oxidation, charring and dry rot—which make minor modifications in the ultimate product.

Concretions.—Concretions are partly the late results of weathering, but it is not rare to find that concretions furnish fragments to the next overlying beds, and those that do this are diagenetic.

Concretions consist of materials like those of cements (page 283), and they probably form in a similar way. Possibly manganese is more abundant in them than in cements. The rounded forms and concentric

¹ GRAHAM, W. A. P., Experiments on the origin of phosphate deposits, *Econ. Geology*, vol. 20, pp. 319–334, 1925.

structures need explanation. Many of them have a nucleus of fossil matter of some sort. This or the rapidity of growth probably influenced the forms of precipitation. See further readings listed.

The peculiar "cone-in-cone" concretions exhibit a cross-fiber vein structure and appear to have grown by replacement starting along bedding planes. Theories as to the cause of the cones are not very clear, but attribute them to pressure which may be (1) load, (2) crystal growth, or (3) from the change of aragonite to calcite.

Reworking and Structural Modifications.—Here are listed a somewhat miscellaneous group of features indicating diagenetic action rather than an outline of principles. Possibly ripple marks from oscillating waves are to be considered diagenetic. The growth of concretions may crowd aside the primary beds and change the structures. If one may judge from the flat pebbles in intraformational conglomerates, wave action has abraded some pebbles in those beds. A crumpled structure of very complex nature may be noted in many shales in a series of beds otherwise almost undeformed. These may result from the slipping of inclined mud deposits down a subaqueous slope. Again, it may be that they grew as the "mud lumps" of the Mississippi delta grow. These appear to be accumulations of somewhat gelatinous clay, probably crowded into regions of light load by an extra load in some other region.¹ The lumps usually have some gassy mud springs also. Borings show much less sand in the lump than at a distance.

ALTERNATIVE ORIGINS OF SOME MINERALOGIC TYPES OF SEDIMENT

A glance over the outline of the petrography of sediments shows several mineralogic types of sediment that may have one of two or more origins. Precipitation from solution may occur either chemically or organically and result in minerals that are hardly distinguishable. Commonly the organic deposit may retain some organic structures, but they may be difficult to recognize. The trend of opinion in recent years seems to be toward the conclusion that if a reaction may be effected either chemically or organically, the *organisms do it more quickly and more completely than the inorganic reagents*. Limestones, iron ores, and cherts are good examples. Many limestones have fossils in a calcareous mud-like groundmass. It may be doubtful whether the matrix is a mechanical mud or a bacterial precipitate. In limonites the abundant surface beds are filled with organic structures, but some deeper deposits are certainly entirely inorganic. Black flints probably owe their color to organic matter from the organism that precipitated the silica, but some late chert concretions have not such abundant fossils as diagenetic cherts and may be wholly inorganic.

¹ SHAW, E. W., The mud lumps at the mouths of the Mississippi, U. S. Geol. Survey Prof. Paper 85, pp. 11-27, 1914.

SEQUENCES AND CYCLES OF SEDIMENTATION

Any broad careful study of sediments will reveal a conspicuous tendency toward cycles of deposition.

Varves.—The clays of many regions show a varved character (page 271). They form a series of thin beds each grading from a coarse-grained base to a fine-grained top and overlain by a later bed that has an exactly similar gradation; so that the contact between beds is marked by an abrupt change from the fine-grained top of one bed to the coarse-grained base of the next (Fig. 164).

Varying Varves.—The thicknesses of such beds in a series are not uniform, and in large exposures a certain amount of rhythm may be detected in the recurrence of the thick beds.

Glacial Sheets.—On a much larger scale an alternation of deposits can be detected in such material as glacial drift. In many parts of North America the drift consists of a succession of three or more sheets separated by surfaces of weathering indicating non-deposition.

Normal Geosynclines.—A much more important feature of sediments is not so readily observed in a single district as in a study of the geologic column through a larger region. For example, if the columnar sections given in a group of folios of the U. S. Geological Survey for a region in the Appalachian Mountains are compiled into a generalized section, it is evident that the sequence of deposition in any period in such a geosyncline is normally

4. Limestone.
3. Shale.
2. Sandstone.
1. Conglomerate (not universally present).

The sequence is in some places on a larger scale than in others (Fig. 213).¹

(See the list of further readings on cycles of sedimentation.)

There are, to be sure, many rocks deposited in the reverse order, and alternations of two rocks many times repeated.

¹ For a general sequence in Mississippian and Pennsylvanian see U. S. Geol. Survey Folio 82. For subordinate sequences in the several divisions of these same series see Folio 133. In Folio 119 the sequences seem related to the sediments between two unconformities but not to the exact divisions between periods.

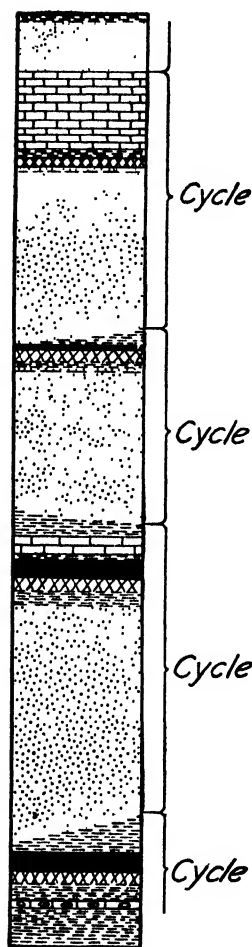


FIG. 213.—Sketch of a geologic column in Illinois, showing several cycles of sedimentation. (After J. M. Weller, *Jour. Geology*, vol. 38, pp. 97–135, 1930.)

Some portions of a cyclic series may be absent, and extra formations may be present. These irregularities usually prove to be minor incidents in the major sequence.

Larger Rhythms.—A rhythm of still larger magnitude is indicated by the recurrence at certain periods in geologic history of almost world-wide deposition of glacial drift; world-wide deposition of coal; world-wide deposition of "red beds" and world-wide deposition of iron ores.

The Nature of the Sedimentary Rhythm.

In the vertical sequence in certain formations the change in sediments is a gradual transition in both directions, but in many more formations there is an abrupt change in one direction followed by a gradual transition back. This manner of occurrence has given rise to some speculations. It is to be expected that slow gradual changes in the conditions of sedimentation will produce only gradual changes in the material deposited, and so the abrupt changes are commonly attributed to catastrophic changes of conditions either orogenic or climatic. A sudden rise of mountains or a sudden increase in rainfall will account for an abrupt increase in coarse mechanical sediment. In a minor way the varved clays also indicate a catastrophic form of rhythm. The summer floods from melting glaciers bring to glacial lakes a great volume of mixed sand and clay; the sand settles first and grades into the clay later during the winter season and before the next floods.

This catastrophic notion of abrupt changes in the sedimentary record is not at all necessary, however, to explain the large cycles. Slow changes producing a broken record may be illustrated by the vertical oscillation of the continental shelf. If the movements changed the depth of water only a little, and were slow, the gradation would follow a smooth curve. But if the slow upward movement once raised the basin floor above sea level, erosion would intervene before the deposition of shallow-water sediments again began, marking a new cycle of deposition. Thus at one stage of a cycle the processes may actually be reversed, and the beginning of the positive part of the cycle may seem abrupt in the sedimentary record (Fig. 214). Corresponding to erosion in mechanical sediments,

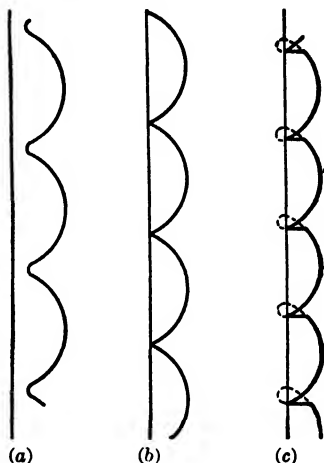


FIG. 214.—Diagrams of the varying rates of sedimentation. (After Barrell, *Geol. Soc. America Bull.*, vol. 28, p. 790, 1917.) Curves *a* and *b* show that deposition may decline to a very slow rate. Curve *c* shows a retrograde portion in each cycle, at which time there is erosion or solution instead of deposition. After such a series of deposits, the next series appears to have started abruptly, but is abrupt only because part of the record is lost. The change of conditions may be abrupt, but is not necessarily so in every case of abrupt change in the sediment.

the retrograde portion of a cycle in chemical and organic deposits is a period of solution. The minor intervals—diastems¹—are likely to be marked by intraformational conglomerates. The portions of formations eroded or dissolved in the negative portion of a cycle are, of course, deposited elsewhere, but they are seldom recognized. It is evident that a bed may form fast, but a formation much more slowly.

Earth Movements and Sedimentation.—A study of the geologic record (page 342) shows a very definite sequence of sediments in many periods—(1) conglomerate, (2) sandstone, (3) shale, (4) limestone. This relation is so world-wide and so generally recognized that it forms the basis of most discussion of sedimentation.

Many features of the geologic record, especially the unconformities at the base of these sequences, suggest a large degree of control of sedimentation by the adjustments of the crust. Questions may arise as to whether there are minor earth movements that determine minor variations in the deposits, and whether those long-separated periods of extreme conditions, such as produced glaciation or widely distributed red beds, can be explained by broad earth movements. This can not be answered so definitely. The indirect effects of changes in the earth's crust can not yet be estimated with much accuracy—changes in air and ocean currents, changes in rain and snowfall, changes in sea level and in snow line, changes in the proportion of land in high latitudes. The forces back of orogeny and isostatic adjustments were considered at length in connection with magmas (pages 159 to 174).

The succession of sediments in a marine series naturally varies with the stability of the basin of deposition. If the crust is stationary, so that the basin is filled as the land is base leveled, the sediments may be expected to grade in the normal order from sandstone to limestone. If the basin is slowly sinking, the same succession is expected but with thicker formations. Sandstones of shallow marine deposition have accumulated to thicknesses of tens of thousands of feet, and have been followed by thick shales and limestones in so many places as to indicate that the normal basin of deposition has a slowly sinking floor. If the sea bottom is rising, the sequence is reversed, and limestones may grade upward to shales and sands, but this sequence is relatively rare. The oscillation of sea bottom in minor adjustments is sufficient to explain the minor alternations in certain members of the general sequence.

Signs That Climate Affects Sedimentation.—The sedimentary effects of climate are equally obvious. Deposits of wind-blown material and salt and gypsum in an arid region contrast with the muds and peat of a moist, cool region. The drift from polar ice sheets is in great contrast to the laterites of the tropics.

Sediments now forming in different climatic districts of the earth show differences analogous to those seen in the older parts of the geologic column. A particularly good example is the red beds, such as are now

¹ BARRELL, JOSEPH, Rhythms and the measurements of geologic time, *Geol. Soc. America Bull.*, vol. 28, p. 794, 1917.

forming in small areas, chiefly where the climate is arid. These beds formed very widely over the earth at three distinct periods in the past.

The rhythmic advance and retreat of the ice in Pleistocene time can hardly be explained except as a climatic effect, for the ice came to sea level in regions that are now near sea level, indicating that there was no important orogenic action.

In Recent time there have been climatic changes of noteworthy importance. Lakes have been dessicated; and in places where there is now very little rainfall there are alluvial fans and records of a civilized habitation, indicating much more rainfall not very long ago. It is said that the growth rings of trees in certain regions and possibly other weather records show a rhythm.

The criteria of certain climatic conditions are here listed, and it should be noted that the signs of rain in past geologic periods may be about as uncertain as the predictions of rain in modern daily life. (See also the readings listed.)

CRITERIA IN SEDIMENTARY ROCKS SUGGESTING CLIMATIC CONDITIONS OF FORMATION

Arid climate with desert conditions—This is indicated by

Wind-blown sands.

Widespread dune structures.

Desert polish and varnish, "case-hardened" boulders and bed-rock, dreikanter and einkanter.

Presence of soluble salts—salt, gypsum, anhydrite, nitrates, borates, etc. Silica and iron oxide concentrate at the surface.

Red color due to iron oxide coating.

Crystals of salt and their casts.

General absence of fossils.

Frosted crystals and fragments of gypsum.

Very fresh feldspar, and arkose.

Semiarid climate.

Abundant mud cracks, footprints, clay galls.

Mild, equable climate, with plenty of moisture.

Muds with high kaolin content, associated with quartz sands and quartz pebble conglomerates. Also a lack of soluble minerals or easily weathered minerals in same.

Lateritic deposits. If humus abundant, silica concentrates at surface; if humus low, alumina and iron concentrate at surface.

Coal seams.

Fossil evidence.

Solution cavities.

Cold climate.

Any evidence of glacial activity, boulder clay, striated and faceted pebbles, etc. Fossil evidence.

Arkose and feldspathic sandstone.

Cracks formed in minerals and rocks by frost and severe temperature changes.

NOTES.

Recent changes may be evident in topographic features.

Seasonal bands, annual layers, etc.—These give evidence of seasonal variations. They may appear in fossil wood, glacial lake clays, halite-anhydrite beds in salt deposits. Variations in the bands from year to year may give indications of climatic cycles.

Glacial and interglacial formations may be interbedded.

Fossils in interglacial beds often give evidence of the extremes in variation of the climate.

Loess is suggestive of arid conditions following a glacial period or more or less contemporaneous with an ice invasion.

Presence of decomposable minerals such as hornblende, augite and biotite in a fine-grained sediment suggests a cold or an arid climate, more probably the former.

Fossil distribution is perhaps the most useful criterion in regard to temperature.

Evidence from corals and some other marine organisms, but particularly from land plants and animals, is fairly definite.

Wet and dry seasons, in absence of humus, give good laterite of Al and Fe but with rich humus, no transport of Al, Fe or Si.

Amount Climate Has Changed.—The change from glacial to hot, arid conditions may not be wholly due to heat, but heat is a most important factor. Although a low temperature with even, moderate precipitation might develop glaciers, a high temperature would dry the area up. It is estimated that the change from winter to summer in our temperate zone is more than enough to explain climates of the geologic past. Imagine 10,000 years like winter at a latitude of 45°, and it is easy to understand the continental glaciers. On the other hand, there are now some areas in which red beds are forming, and a little more heat would no doubt extend the areas considerably. Estimates range from 10° to 40°F. for the change in mean annual temperature needed to account for the changes in the geologic past. These are not great changes.

Basis of Climatic Changes.—Changes in geologic climates are much discussed, and many suggestions seem to be quite inapplicable to explaining the sedimentary record.

Certain changes in the atmosphere no doubt affect climate. Dust from volcanoes or from nebulae might affect heat transfers, but no relations between climate and volcanic explosions can be detected in the geologic record. It is found also that H₂O and CO₂ in the atmosphere exert a blanketing influence, and it is known that in certain periods the removal of CO₂ from the air must have been very great to form the beds of coal and limestone. Nevertheless, a close study indicates that the CO₂ now present does about all the blanketing that can be done; and the ocean is such a vast reservoir holding CO₂ in solution and in equilibrium with the air that the change would have to be very great to have a notable effect.

The relation of climate to elevation is well recognized. The higher regions are cold, and the winds that come from an ocean to a mountain slope lose much of their moisture as rain. The rise of mountains may change the climate not only of the region elevated but also of all that region swept by winds that have crossed the mountains.

It is also probable that ridges may somewhat deflect both air and water currents. Although it may be admitted that most winds are planetary and that most ocean currents are determined by winds, it is not certain that the present system has always prevailed. A barrier raised across some ocean current would have effects that are hard to predict.

And there are two causes of oceanic density variations, temperature and concentration. These causes are somewhat opposed, and at some geologic time the currents may have been reversed.

Note should also be made of some of the ideas of Joly and of Holmes that the earth has a substratum that is periodically heated to fusion by the heat of atomic disintegration, and that continents may drift on the molten zone. The molten layer thus comes close below the ocean and gives up some of its heat to the ocean (see page 182). The rise in temperature may not be great but might indirectly have large climatic effects by increasing the evaporation and rainfall. The drifting of the continents might also shift the axis of rotation of the earth.

Solar radiation is now being carefully observed, and it is safe to say that an important variation takes place. The variation is said to be related to sun spots, an increase of 100 in the sun-spot number marking an increase of heat of perhaps 4 per cent. Many attempts are also made to correlate the sun-spot record and other climatic records, both historic and archeologic. The growth rings of trees are said to match the sun-spot cycle of $11\frac{1}{2}$ years for 400 years back (see page 206), and the sequoias show traces of such a cycle even back to 1300 B.C. The triple cycle of about 35 years is also marked, and there may be longer cycles. Some stars are variable (in long, short, or erratic periods) in color and brightness and probably also in heat radiated, so that our sun by analogy probably varies in heat radiated. Attempts are made to extrapolate from these solar cycles to the glacial periods of the Pleistocene, but although the idea is not illogical, the evidences are slight. If reduced solar radiation caused continental glaciation, the polar ice caps should have widened for the whole earth. On the contrary, Pleistocene ice sheets in North America progressed from Cordilleran to Keewatin to Labradorean. Peruvian glaciation was similarly local. De Geer, however, concluded from a study of varved clays that Pleistocene glaciations were isochronic and had a common cause.¹

It should be noted that some slight temperature changes, whether orogenic or solar, may have local effects of great moment because the whole system of cyclonic storms is shifted by them. If the solar radiation changed one per cent for a few years, it might change the mean annual temperature at certain points in North America as much as 40°F. and increase rainfall 100 per cent.

Summary.—General conclusions as to the causes of sedimentary rhythms are not satisfactory. No one cause seems to fit all cases. Some broad variations are probably attributable to climatic changes from solar radiation, others almost world-wide are attributable to orogeny. Local effects may be orogenic or due to shifting currents of air and water. Some merely local effects, however, seem to be unrelated to any local

¹ DE GEER, G., Late glacial clay varves in Argentina, *Geog. Annals*, vols. 141-142, p. 7, Stockholm, 1927.

changes of level. The recent suggestion of periodic drifting of continents is attractive, but needs considerable further testing before it is acceptable for many specific cases in the sedimentary record.

INTERPRETATIVE PETROGRAPHY OF SEDIMENTS

The interpretation of sedimentary rock features in terms of the source, the conditions of transportation and the places and conditions of deposition is a relatively new subject (see the readings listed), and the principles now tentatively agreed upon may need revision or restatement after further study. All the remarks and theories in this section may be involved in the interpretation (but see especially the tabulated criteria, pages 310, 311, 313, 316, 322, 334 and 345).

The problem of source of sediments is attacked partly by broad structural and paleogeographic studies but largely also from the petrographic side. Certain studies of cross-bedding due to currents indicate the direction from which sediments came. More definitely petrographically, the heavy residual minerals may prove to be of such kinds or to have such peculiarities of form, color, etc., as to indicate a certain igneous or metamorphic rock as the most probable source of supply. With due caution this method may be of great value; but the caution is especially needed where exposures of the source rocks are now largely inaccessible. Sediments derived from igneous and metamorphic rocks have more variety in minerals than those formed by the reworking of older sedimentary rocks. As a sedimentary series may have different minerals, especially heavy residuals, in different beds, when such a series is reworked into another sediment, the minerals will occur in the reverse order. The variety and abundance of heavy residuals in a formation may be rejuvenated by tectonic movements of the source rocks.

The conditions of transportation are fairly clear for a few formations, such as the glacial drift; or frosted sands with clearly eolian cross-bedding and dreikanter (see pages 314 to 316); or organic limestones transported in solution, or precipitates like salt and gypsum. In more abundant sandstones a few more obscure features may be suggestive. The variety of the abundant minerals varies inversely as the distance from the source. The variety in heavy residuals increases as the distance of transportation increases and as the number of sources from which contributions are received increases, especially if from igneous and metamorphic rocks; but the total per cent of heavy residuals may decrease. Bands rich in heavy minerals, even though not with much variety, indicate wind and water currents, probably stream or wave action, and are not likely to occur in water deeper than wave action extends.

The conditions and place of deposition are indicated in part by the conclusions as to transportation (see pages 316 and 322).

Certain instances of the application of such interpretations may serve to emphasize their value. A number of commercial deposits occur as sediments or associated with sediments. If they occur with marine deposits laid down in fairly deep water, such as shale, the indication is that at the time of deposition the conditions were uniform over wide areas, and the deposits themselves are likely to be uniform and persistent. Quite on the contrary, if a sediment, such as a coal bed, is associated with sandstones that are cross-bedded and give other evidence of shore conditions and irregularities of deposition and erosion, it must be expected that the coal will show irregularities in thickness and probably also in nature; it may be wholly eroded in many places.

Again, when the petrographic characters of the formations of a region are known, and a series of samples and diagrams are available of texture, shape, mineral suites, and possibly other features, then the characters of an unknown sample can usually be matched to one of the known formations better than to any other. The distance over which a formation remains fairly constant can not be stated in general terms. Some vary notably in 5 miles, others may be fairly uniform for 100 miles. Few can be correlated by these methods for great distances, and the method is best for comparatively local use.

Finally, it may be noted that an increase of sand from one place to another in petroleum-bearing beds suggests an approach to shore conditions and possibly lenses of porous sand in shale. (See the list of further readings.)

PART VI

THE PETROGRAPHY OF METAMORPHIC ROCKS

GENERAL FEATURES

Metamorphism in rocks may range from slight to extremely great. Rocks in which the changes are slight are not commonly classified as metamorphic. Metamorphic rocks are those whose *distinctive characters* are produced by metamorphism. There is considerable difference in the judgment of different authorities as to how great a change makes a rock metamorphic (see the readings listed).

The essential change in a metamorphic rock is new crystallization or new texture throughout the body of the rock. Even this is subject to considerable interpretation, and it may be necessary to define by a series of illustrations. An obsidian that has been devitrified is still commonly classed as igneous, unless it has been deformed. Most petrographers exclude surface-weathered products from metamorphic rocks. A clay no doubt develops new crystallization by the time it becomes indurated as shale, but it is rarely classed as metamorphic unless it becomes coarse or slaty. Very different is a sandstone in which the cementation is a result of secondary growth (new crystallization) of the quartz grains. If the cement is so firm that the rock breaks across the grains as easily as around them, the quartzite is called a metamorphic rock.

The general characteristics of metamorphic rocks include a banded structure and a texture that involves intergrown interlocking crystals (see the Criteria, pages 7 and 8).

METHODS FOR METAMORPHIC PETROGRAPHY

The description of metamorphic rocks involves little that is new after igneous and sedimentary rocks have been studied. Certain structures and minerals are characteristic, and the object of the study is very commonly to determine whether the metamorphic rock was originally a sediment or an igneous rock.

FIELD SCHEDULE FOR EXAMINING METAMORPHIC ROCKS

1. Location.
2. Name.
3. Form and boundary characters.
Related igneous rocks, *lit-par-lit* or sharp contacts.
4. Schistosity, dip and strike, relation of drag folds.
5. Bedding, dip and strike and how identified; pitch of folds and flow lines.
6. Joints and other structures.
7. Textures.
8. Minerals.
9. Variations.
Collection of samples and specimens.

METAMORPHIC STRUCTURES AND TEXTURES

A large proportion of metamorphic rocks are foliated. A *foliate* is a rock which exhibits a parallelism of certain or all of its mineral constituents. The parallelism may be in the elongation of single grains or in the elongation of bands, layers or lenses of granular material (Fig. 215). In the latter case a foliation may be distinguished with difficulty from lamination or bedding in sediments, but the term foliation is applied mostly to rocks with an intergrowth or interlocking of grains, as distinct from the lamination of deposits consisting of fragments more or less cemented by other materials. Foliated rocks are thus chiefly metamorphic; they comprise the gneisses, schists and slates but include a considerable number of igneous rocks that show a flow structure, trachitic or trachitoid. Metamorphic rocks are chiefly foliated, but a few are massive.

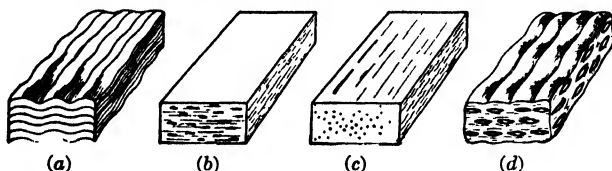


FIG. 215.—Diagrams of the different structures causing schistosity. (a) Crumpled plates. (b) Platy grains. (c) Linear grains. (d) Lenses.

If the parallel crystals are pyroxenes or feldspar, or if the matrix around the parallel crystals is glassy or so fine as to be sharply contrasted in grain size, the rock is probably igneous. The metamorphic rocks do not have glass, and the common parallel minerals in them are biotite, hornblende and chlorite. See the tabulated criteria. The dynamic action that occurs at a late stage of igneous crystallization (frontispiece and Fig. 17) or alternates with igneous injection may leave some rocks in doubt (see piezocrystallization, pages 59 and 440).

Metamorphic rocks commonly have a cleavage, and two kinds may be distinguished, "fracture cleavage" and "flow cleavage" (see pages 427 to 429).

CRITERIA FOR DISTINGUISHING METAMORPHIC AND PRIMARY IGNEOUS FOLIATION

These criteria, largely from Leith,¹ are complicated by the fact that foliated rocks include injection gneisses, in which the foliation is metamorphic, but much of the material is igneous. The rocks are foliated because the schistosity guided the injection. Such rocks are not clearly covered in this tabulation (see frontispiece and Fig. 223).

A. Primary foliation in igneous rocks.

Field.—Foliated on all the margins, roof and walls, but massive in the center. An intrusive with foliation parallel to the walls but at an angle to foliation in the wall rock or between walls that have no foliation.

Foreign inclusions with less foliation or a different foliation from that of the matrix. Distinct bands of different proportions of the same minerals where not sheared or injected.

¹LEITH, C. K., "Structural Geology," rev. ed., p. 231, 1923.

Absence of sharp contact between silicic and basic folia.

Bands may curve or even close in a circle.

Associated pegmatites that seem clearly related.

Textural.—The order of crystallization of an igneous rock, not that of the crystalloblastic series; usually granitoid.

The needles and platy minerals always flow out around any lumps like phenocrysts or inclusions.

Any glass, glassy matrix, diabase, or other clearly igneous texture (see criteria, pages 7 to 8).

Minerals.—If all the minerals are igneous, foliation is likely to be. Especially if nephelinite or olivine or minerals easily altered by metamorphism. The foliation due to feldspars or pyroxene is often igneous; that from hornblende in euhedral grains, also probably igneous.

B. Metamorphic foliation.

Field.—Any sign that the rock was sedimentary (see criteria of origin of metamorphic rocks, page 431).

Foliation of the borders of an igneous mass by metamorphism is likely to be patchy and chiefly where the wall rock is especially deformed. It may be parallel to the border but in some intrusions is independent of that direction and more nearly parallel to that in the walls.

Field signs of deformation.

Bands have sharply contrasted minerals rather than varying proportions of the same minerals.

Textural.—Crystalloblastic series (page 356).

Strain effects in many grains.

Metacrysts, augen structures, mortar structure, mylonite.

If the schistosity is transgressed by metacrysts, it is metamorphic. If the schistosity bends out around the large crystal, the origin is best judged from the mineral of the large crystal.

Helizitic structure, metacrysts with oriented inclusions of earlier schist minerals.

Minerals.—Those rare except in metamorphic rocks or not found in igneous rocks; staurolite, sillimanite, kyanite, cordierite, etc.

The foliation is due largely to biotite and chlorite; hornblende in some rocks (see page 357).

The following terms cover most of the textures and structures of metamorphic rocks:

Slaty. Foliated, without parallel lamination, fine grain (page 365).

Schistose. Foliated, usually without lamination, medium grain (page 362).

Gneissic. Banded, streaked, coarse grain (see frontispiece and page 358).

Relict or *palimpsest* structures. Relics of original rock in a rock that has new metamorphic structure.

F. Becke¹ has introduced a series of terms based on the root word blast, to sprout. As a prefix it indicates original (relict) textures, as a suffix the newly formed textures (Figs. 217 and 218).

Crystalloblastic structure is described as covering several points, chiefly that in most cases all constituents lack crystal form, but if crystal forms occur they are simple. Poikilitic, sometimes oriented, inclusions are common and zoned structures are rare. Glass is not found.

Granoblastic = mosaic = granulitic = hornfels = sugary grained.

¹ *Über Mineralbestand und Struktur der kristallinen Schiefer*, Cong. géol. internat., 9th sess., pp. 553 et seq., Vienna, 1903.

Porphyroblastic = maculose = having metacrysts, large crystals formed by recrystallization in a finer matrix of schist or slate (Figs. 216 and 261).

Poikiloblastic, having large crystals with many large inclusions.

Sutured. With complex interlocking boundaries (Fig. 219b).

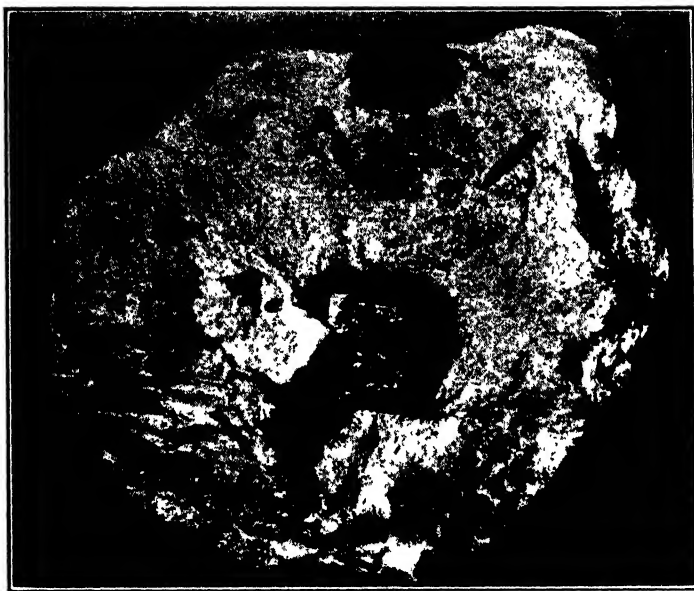


FIG. 216.—Porphyroblastic texture. Garnet metacrysts in a muscovite schist, Roxbury, Connecticut. About natural size.



FIG. 217.—Photomicrograph of a graphitic slate with cordierite metacrysts. Porphyroblastic texture to compare with Fig. 218. $\times 20$.

Mortar structure. A little crushing between grains, resulting in an aggregate like stones in a wall with mortar between (Fig. 219d).

Augen structure. Lens of some mineral in a crushed matrix (Figs. 219e and 220).

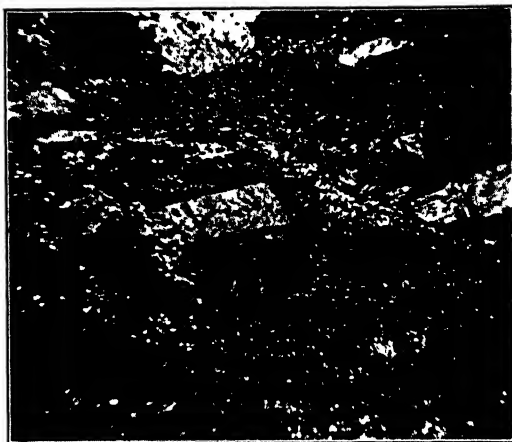
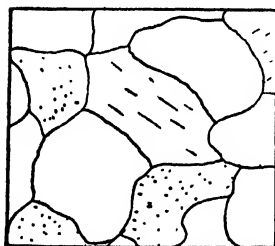
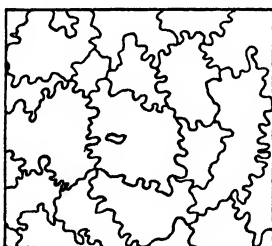


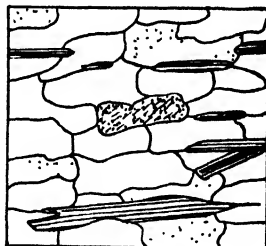
FIG 218.—Photomicrograph of a schistose rhyolite porphyry, Porcupine, Ontario. Blastoporphyritic texture, to compare with Figs. 216 and 217. Crossed nicols. One Carlsbad twin of orthoclase is broken and the fragments separated. Most of the original phenocrysts are obscured by recrystallization and deformation. $\times 33$.



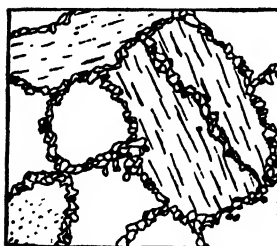
(a) Granoblastic or granitoid (mosaic).



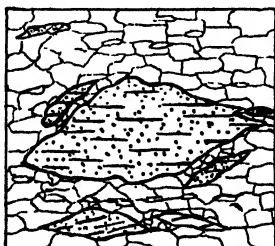
(b) Sutured.



(c) Recrystallized under stress (schistose).



(d) Mortar structure.



(e) Augen structure.



(f) Flaser structure.

FIG. 219.—Diagram showing the appearance in thin sections of the several textures commonly resulting from progressive granulation and recrystallization.

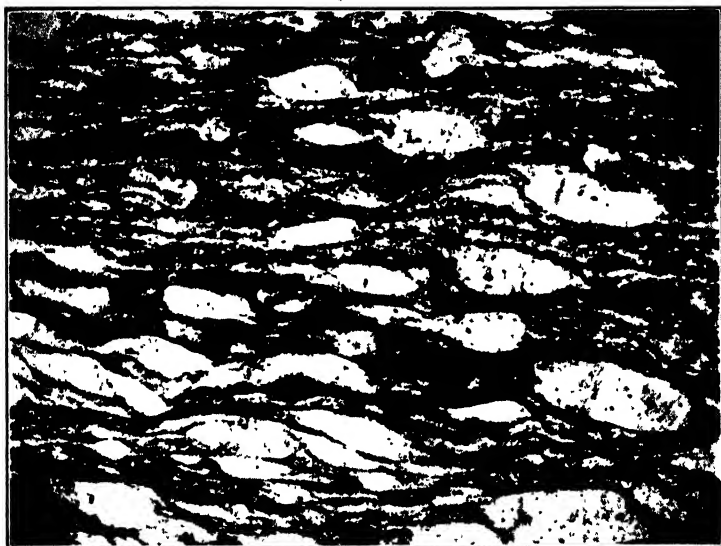


FIG. 220.—Hand specimen of augen gneiss, South America. About one-half natural size.



FIG. 221.—Photomicrograph of a thin section of graywacke that has been crushed so that one plagioclase grain now shows in the section over 20 fragments, but none that have moved far. Mylonite. Crossed nicols. $\times 12$.

Flaser structure. Lenses of a mineral complex or rock in a crushed or sheared matrix (Fig. 219f).

Mylonitic structure. Crushed completely (Fig. 221).

Folded, faulted, sliced, sheared, shattered, healed, veined, replaced, pseudomorphous, radiating, feathered, with secondary growth, granulated.

CRITERIA FOR DISTINGUISHING METACRYSTS AND PHENOCRYSTS

(See also page 281, criteria of fragments and amygdules and phenocrysts)

Minerals.

Strongly indicating metacrysts are garnet, staurolite, ottrelite, kyanite, andalusite, cordierite, pyrite.

Uncertain are magnetite, ilmenite, tourmaline, biotite, hornblende, augite, feldspars, quartz, olivine.

Matrix.

Glass or characteristic igneous groundmass—phenocryst; argillite or characteristic sedimentary material—metacrysts. If the matrix has a flow structure, see pages 351 and 352; if the large crystal transgresses the structure—metacryst. If structure bends around an equidimensional large crystal it is still uncertain. If the structure bends around an elongated large crystal, the large phenocrysts are oriented with the matrix, but the metacrysts are independent of it; so if the crystals transgress the structure they are metacrysts, if parallel they are still uncertain. Associated amygdules suggest phenocrysts.

Inclusions.

More abundant and coarser in metacrysts; oriented inclusions indicate metacrysts. Corrosion of large crystals always phenocrysts. Granulation leaves it uncertain.

MINERALS OF METAMORPHIC ROCKS

Practically any rock mineral may be found at some place in a metamorphic rock. Probably the dominant minerals are quartz, carbonates, micas, feldspars, hornblende, chlorite, epidote, magnetite, and pyroxenes. Certain kinds of rocks may have restricted lists, but generalizations have relatively little value. As the rocks formed by weathering are classified as sediments, the minerals characteristic of complete weathering, kaolin, limonite, bauxite, etc., are not commonly present in the metamorphosed sediments. Orthogneisses have few minerals aside from those listed as characteristic of igneous rock clans.

The following minerals are *nearly always metamorphic*; those marked with an asterisk are common in contact metamorphic rocks: staurolite, garnet*, andalusite*, sillimanite, kyanite, glaucophane, talc, cordierite*, wollastonite*, vesuvianite*, brittle mica, graphite*, tremolite, chondrodite*, zoisite*, anatase, rutile, leucoxene, actinolite, anthophyllite, scapolite*, phlogopite*. (See also the criteria of the three main classes of rocks, pages 7 and 8.)

THE CRYSTALLOBLASTIC SERIES

The order in which minerals develop euhedral crystals during recrystallization can not well be estimated by their textural relations. The usual

series based on form development is somewhat as follows,¹ with the early-named minerals of the list most inclined to be euhedral: titanite, rutile, hematite, ilmenite, pyrite, garnet, tourmaline, staurolite, kyanite, epidote, zoisite, pyroxene, hornblende, magnetite, dolomite, albite, mica, chlorite, talc, calcite, quartz, plagioclase, orthoclase, microcline. In general, this series of minerals shows a decreasing specific gravity from first to last, but the significance of that fact is uncertain. The order probably depends on crystal forces.

There is, furthermore, probably a mineral sequence in the assumption of parallelism during recrystallization under stress. Calcite easily shears into parallel plates but also recrystallizes easily to obliterate that structure. Sericite develops early from either feldspar or clay; other micas, chlorite and hornblende precede quartz, and quartz may precede feldspar. Further metamorphism produces parallelism in such minerals as tourmaline, sillimanite and other amphiboles.

There is no doubt a sequence also in the development of new minerals during deformation, but as the metamorphism involves a cycle, the early stages of the cycle may make no record but that which is obliterated by the later action. For example, metacrysts may form after a group of platy minerals in a schist, but there may also have been metacrysts before the deformation that developed platy minerals (see page 437 and Fig. 261).

Besides producing new minerals, textures and structures, metamorphism often produces some internal effects in the mineral grains—bent crystals, strain shadows or wavy extinction, and a pressure twinning. It is difficult, however, to distinguish these effects of metamorphism from similar internal features that seem to be independent of strain. Flamboyant quartz in a vein has a wavy extinction much like that in deformed quartzites. Calcite in nearly all rocks has twinning such as is developed in deformed limestones. The microcline of pegmatites may show twinning much like that developed in orthoclase by deformation. Nevertheless along with other criteria these features indicate dynamic action (see criteria, pages 376 and 411).

CLASSIFICATION OF METAMORPHIC ROCKS

Some elaborate schemes of classification have been proposed, and in advanced work and detailed study these may be useful (see the list of further readings). Nearly all workers agree to use variety names based on terms indicating minerals and textures, prefixed to the common group names used in elementary work. Only the prefix needs discussion. (See also the remarks on pages 139 and 140 under the heading *Alteration of igneous rocks* and on page 336 under *Diagenesis of sediments*.)

The suggestions for terminology are in great confusion. *Meta*—prefixed to a rock name of some other class indicates it has been metamorphosed, but Lacroix proposes restricting it to those chemically altered, and Rosenbusch used it for rocks of unknown original. *Katamorphism* is Van Hise's term for alterations near the *surface*, but Grubenmann prefixes *kata-* to rock names to indicate the *deepest* of three zones of metamorphism. Lacroix prefixes *epi-* to rock names to indicate that the rocks

¹ GRUBENMANN, ULRICH, "Die Kristallinen Schiefer," p. 74, 1904.

are of uncertain original, but Grubenmann uses the same prefix for rocks altered in the uppermost of three zones regardless of origin. Para- was used by Rosenbusch to indicate a sedimentary original, but by Lacroix to show that minerals had changed without much chemical change in the rock. Probably none of the suggested prefixes is to be recommended for general use. (See the readings listed.)

GNEISSES

Definition.—Gneiss is a megascopically crystalline rock with a secondary rough foliation developed as a result of pressure on the solidified rock; the bands or folia or lenses in gneisses are commonly unlike. The mineral grains may or may not be elongated and arranged (see frontispiece



FIG. 222.—The surface of a hand specimen of diorite gneiss, Mt. Diablo, California, showing grains oriented by metamorphism, but no prominent banding. About natural size.

and Figs. 222 and 223). In a large majority of gneisses feldspar is a prominent constituent, and it is customary to let abundant feldspar distinguish a gneiss from a schist.

This definition excludes the trachitoid igneous rocks sometimes called "primary gneiss" but includes those igneous rocks that have been crushed or sheared into bands and those schists that have been injected by igneous magma *lit-par-lit*.

A. Textural and structural terms.**Granoblastic, etc.** (page 352).**Augen gneiss.****Mortar structured gneiss.****Sheared gneiss.****Mylonites** (thoroughly granulated but with streaked minerals).**Pseudotachylite** (crush-rocks of almost glassy fineness of grain).

FIG. 223.—An outcrop of banded gneiss, Saganaga Lake, Ontario. The dark rock is schistose and possibly of sedimentary origin, but it grades by the injection of magma into a gneiss of dominantly igneous material:

B. Names based on original class of rock involved and process of change.**Orthogneisses**, from igneous originals.**Paragneiss**, from sedimentary originals.

If the rock is of unknown origin, as many are, it needs no prefix, but the prefix "meta" can be used.

Injection gneisses (migmatites), from schists of any origin after injection of igneous rock *lit-par-lit*.

Conglomerate gneiss.**Quartzite gneiss.**(See also *C* below.)**C. Names indicating composition.**

1. Based on the names of granitoid igneous rocks of the same mineral composition,¹ as, for examples, granite gneiss and gabbro gneiss.)

¹ It has been suggested, however, that the distinctions on original class of rock might be shown by varying the qualifier. For example in the granite family:

"Granite gneiss," the orthogneiss.

"Quartz orthoclase paragneiss," the altered sediment.

"Granitic gneiss," the rock of unknown origin.

"Injected granite gneiss," the rock of mixed origin.

In case of uncertainty the name would be "granitic gneiss" and probably would give a fairly definite notion of the rock to most geologists. Similar terms are available in other families.

2. (Based on prominent minerals. Few of these are as completely descriptive as the names based on the corresponding igneous rock, hence much less satisfactory.)

D. Special terms.

Eclogite, garnet and pyroxene, etc.

Granulite, garnet, pyroxene, and feldspars, etc.

Skarn, a Scandinavian term for gangue or country rock of contact magnetite ores.

Protogine, old granites and gneisses, nearly all sheared and probably originally trachitoid.

Leptite, mostly recrystallized silicic aphanites, may be porphyritic or even-grained.

Finer than most gneisses but not very schistose.

"**Flaser**"—as a prefix to the names of granitoid igneous rocks indicates a sheared condition.

Notes on Gneisses.—The banding of gneisses is typically an alternation of layers which may or may not show a parallelism of grain. Bands

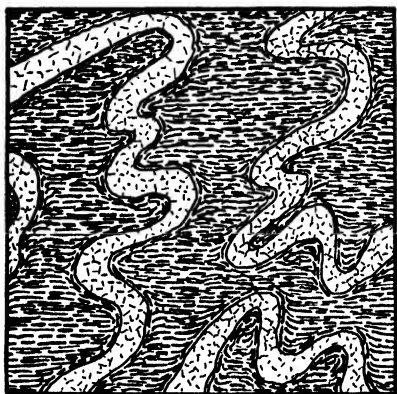


FIG. 224.—Pygmy folds in schist.
About natural size.

may be conspicuously curved or irregular. Some gneisses are very little banded but do show parallelism of grain. The evident differences in origin of different gneisses are not all distinguished by definite criteria. Besides the "primary gneisses," which are igneous, one has to distinguish injection gneisses (Fig. 223) and those that are formed by deformation of an igneous rock at the late stage of magmatic action, also those of distinctly later metamorphism (see pages 351 and 431). The injection of foliated rocks during deformation

probably accounts for "pygmy folds" (Fig. 224).

Mortar structure, augen structure and mylonite indicate progressive granulation by crushing. Many micas and other flexible minerals show bending rather than granulation.

Granulites often have granular fringes of other minerals arranged around the borders of rounded garnet grains, but the garnets seldom form metacrysts in the gneisses, as they do in schists.

Eclogite is much like basalt in composition but contains other and heavier minerals. It may be regarded as a deep-seated equivalent of rocks that form basalts near the surface.

Many gneisses have inclusions analogous to the inclusions in granites. There are also in many gneisses related pegmatites and aplites, some of which may have escaped the dynamic action that made a gneiss of the parent rock.

Gneisses may form from granite-pebble conglomerates and still show traces of the original pebbles. If such rocks are thoroughly recrystallized they become "recomposed granite."

Minerals of Gneisses.—The minerals of gneisses are commonly those of granitoid igneous rocks, but the proportions of those minerals may be different from those in common igneous rocks. The feldspars, quartz and micas predominate. Hornblende, garnet, epidote, tourmaline and chlorite are common. Cordierite, graphite and sillimanite are much more common in gneisses than in granites (Fig. 225).

Injection gneisses mix up the mineral composition, adding dikelets that are truly igneous to a schist that may retain some minerals not



FIG. 225.—Photomicrograph of a thin section of gneiss, Lunzenau. The main field is a transparent cordierite, in which are abundant sillimanite needles and three zircon inclusions with conspicuous yellow pleochroic halos. Plain light. $\times 100$.

expected in igneous rock. There is a tendency, however, for injected schist to react with the magma injecting it and form such minerals as are in equilibrium with the magma (see the reaction series, pages 226 and 242). A wormy intergrowth of quartz and albite may also result from injection of magma into feldspar rock; this intergrowth so closely resembles myrmekite (Fig. 35) as to make its origin in the gneisses uncertain. Wormy intergrowths in metamorphic rocks are called symplektite if composed of quartz and feldspar, and kelyphite if composed of other secondary minerals in various combinations. Their metamorphic origin is clear if composed of secondary minerals like calcite or zoisite, and especially if they replace secondary minerals; fairly clear if they show hydrothermal minerals and general recrystallization. They commonly

work into earlier minerals along cleavages in a manner very different from the replacement of feldspar by myrmekite. The feldspars may be of any common color, but it is noteworthy that metamorphosed granites are likely to have gray feldspar, and less-altered granites pink feldspar. The grains in many gneisses are bent or broken (Figs. 218, 221, and 238).

The quartz of gneisses and schists has commonly a pronounced wavy extinction (Fig. 226), which is more banded than that in granites (page 63).



FIG. 226.—Photomicrograph of a thin section of granite gneiss, Lake Athabaska. The main central grain is a quartz crystal showing the strain shadows characteristic of metamorphism. These are much more conspicuous than those in the quartz of unaltered granite and in some grains roughly banded in nearly parallel bands. It may easily be confused with flamboyant structure in vein quartz (Fig. 194). The structure is visible only when the grain is near extinction. Crossed nicols. $\times 100$.

SCHISTS

Definition.—Schist is a megascopically crystalline rock with a secondary foliated or laminated structure; as a rule it is finer grained than gneiss; its foliation is based on parallelism of platy or elongated grains; and it rarely has abundant feldspar (Figs. 227 and 228). Schists commonly split into thinner layers than gneisses (the name refers to the tendency to split). The thin layers are commonly but not necessarily of the same mineral composition as the layers on each side.

A. Textural and structural terms.

Crumpled (or curly) schist.

Phyllite (see slate).

Banded schist.

Porphyroblastic schist (see Fig. 261).

B. Compositional terms. (Prefix the name of minerals that are abundant or conspicuous. Two prominent minerals may be hyphenated as a prefix.)

Grubenmann distinguished by analyses 12 compositional varieties in crystalline metamorphic rocks: (1) orthoclase, (2) aluminum silicates, (3) plagioclase, (4) eclogite-amphibolite, (5) magnesium silicates, (6) jadeite, (7) chlorite-meionite, (8) quartzite, (9) calcium silicate, (10) marbles, (11) iron oxides, (12) aluminum oxides.

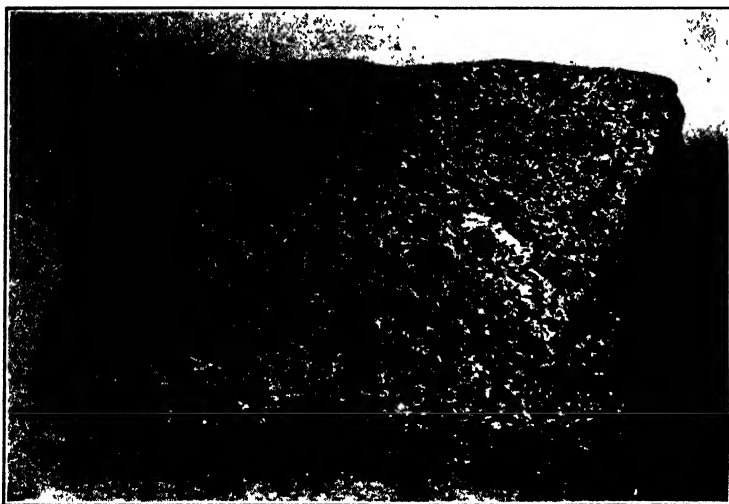


FIG. 227.—Hand specimen of actinolite schist, with linear to platy schistosity. About one-half natural size.

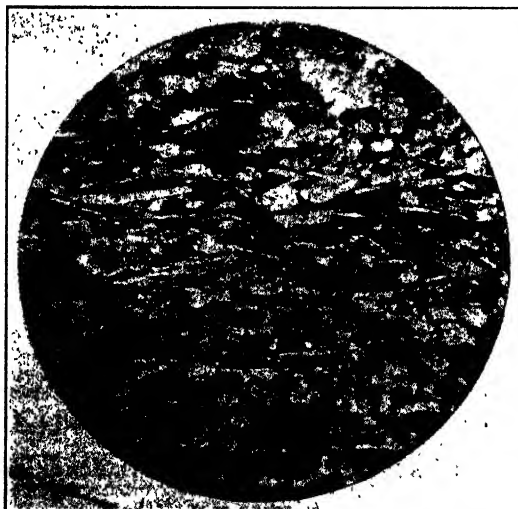


FIG. 228.—Photomicrograph of hornblende schist. Ely, Minnesota. $\times 45$.

C. The zone or facies classification of schists.

It is universally recognized that as metamorphism varies in the intensity of the heat and deformation involved, the mineral assemblages that result from it vary. A facies is a group of rocks characterized by a definite set of minerals, which under the conditions of their formation were in perfect equilibrium with each other. Not enough work has yet been done to determine how many zones or facies it is well to recognize in nomenclature. Clearly the deep zone gives "kata-" rocks with a mineral assemblage different from that of the shallow "epi-" rocks. In a few cases there may be doubt enough to justify an intermediate zone of "meso-" rocks.

D. Special and locality terms.

Leptite (see gneisses).

Greenstone; includes propylite and chlorite schist.

Knotenschiefer, fruchtschiefer.

The distinction of ortho-, para- and meta-schists is similar to that in gneisses, but these prefixes are not much used for schists.

Notes on Schists.—Schists grade into gneisses through several intermediate sorts of rock. Division lines are arbitrary. In the opposite direction they grade through phyllites to slates.

Schists formed by granulation may have here and there lenses or augen of original minerals or textures. Micas are commonly bent; the contacts between minerals may show intergrowths.

Usually the mineral names prefixed to a general term such as schist or gneiss serve very well to indicate the zone or facies, but in some rocks where the mineral assemblage is complex a name for the facies may prove useful. The mineralogic distinctions are commonly accompanied by differences in structure; epi-schists show little deformation of pebbles, amygdulæ and other original structures and may be notably fractured but not much recrystallized; kata-schists are wholly recrystallized and may have entirely new minerals and structures.

Eskola suggested as facies with increasing degrees, or grades or intensities of metamorphism (1) hornfels facies, (2) sanidinite facies, (3) green schist facies, (4) amphibolite facies, (5) eclogite facies. Each of these may have mineralogic qualifiers. Thus a biotite schist might by some variations in metacrysts or textures prove to belong to the hornfels facies or to the green schist facies. No exact temperature limits have been assigned.

T. Vogt has applied the idea of facies to a double series of metamorphic rocks in Sulitelma, one originally gabbro, the other originally sedimentary. A complete series of facies was found for each rock. He believes the most important mineralogic criteria of temperature are the formation and destruction of (1) anorthite, (2) hornblende, and (3) epidote. These may well be used to mark main divisions and the others to subdivide.

Minerals of Schists.—The following minerals are the basis of the foliations:

Biotite, sericite and other micas; hornblende and other amphiboles; chlorite, chloritoid and other micaceous minerals; talc and serpentine; graphite; pyroxenes; plagioclase; sillimanite; epidote; ilmenite and specularite; tourmaline.

The minerals commonly in metacrysts are:

Garnet; staurolite; ottrelite, chloritoid; magnetite, ilmenite; kyanite; andalusite; tourmaline; cordierite; pyrite; biotite; hornblende; feldspar.

Metacrysts commonly have abundant inclusions (Fig. 229) and in some the inclusions are oriented so as to show that the metacrysts grew by replacement (Fig. 248).

The common minerals of schists are quartz, carbonate, and a little feldspar. These are not commonly coarse grained or notably elongated and are much less conspicuous to the naked eye than their abundance would suggest. These, along with the elongated minerals biotite, hornblende and chlorite, make up all but a very small percentage of the schists. Quartz, carbonate, and pyrite are common additions to schists in veins and pods. The most common garnet in gneisses and schists is almandite.



FIG. 229.—Sketch of thin section of biotite schist with staurolite metacryst enclosing quartz and graphite. Little Falls, Minnesota. $\times 15$.

ARGILLITES AND SLATES

Definitions.—A slate has minerals that are too fine grained to be identified megascopically; it has a tendency to split into thin layers as a result of recrystallization under pressure, and the thin layers are normally like adjacent layers in mineral composition. The surfaces of the split slabs tend to be smoother than those of schist because of the finer grain, but some slates have crumpled and folded cleavage surfaces.

An argillite is recrystallized to about the same extent as a slate but differs in being necessarily a derivative of clay and having no secondary cleavage, only a lamination or bedding.

A. Textural and structural terms.

Curly or crumpled slate.

Bedded slates show sedimentary structures crossed by the cleavage.

Phyllite is coarser and more lustrous than most slates, but still too fine grained to be typical schist.

B. Mineralogic varieties.

(Prefix color terms, chiefly gray, black, red and green.)

Calcareous slate.

Carbonaceous slate.

(And so on as in shales.)

Quartzose slate, may be sandy.

Graywacke slate, grading into graywacke quartzite.
(The names of metacrysts may be prefixed.)

C. Special.

Roofing slate; has notably good cleavage and uniform color.

Minerals of Slates.—With high magnifications some minerals can be detected in argillites and slates. The common minerals are those found in coarser grains in schists, namely, quartz, sericite, biotite, chlorite, graphite, carbonates, and titanium oxides. A great many others occur in smaller amounts or rarer exposures.

Metacrysts are not common; pyrite, magnetite, cordierite and feldspar may serve as examples (see Fig. 230).

Veinlets of quartz and carbonate are very common.

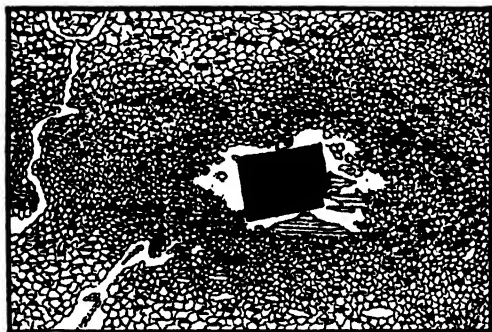


FIG. 230.—Sketch of pyrite metacryst in a thin section of slate, Minnesota. Quartz and chlorite make a sort of halo around the pyrite, chiefly at the ends where the deforming pressure was partly resisted by the pyrite. Some of the structure lines bend around the pyrite and others are truncated by the pyrite and its halo. The pyrite may partly replace and partly displace the slate. (*Econ. Geology*, vol. 21, p. 726, 1926.) $\times 30$.

QUARTZITES

Definition.—A metamorphic rock consisting largely of quartz fragments so thoroughly cemented or recrystallized that the rock breaks through the grains as easily as around them.

A. Textural and structural varieties.

Conglomerate quartzite	}	Indicating size of grain
Quartzite grit		
Fine-grained quartzite		
Novaculite		
Mosaic quartzite	}	Indicating relations of grains
Sutured quartzite		
Quartzite with secondary growth		
Granoblastic quartzite		
Granulated quartzite		

(Augen and mortar structures)

Slaty and schistose quartzites have the texture modified by the impurities present.

B. Compositional terms. (Prefix color terms; chiefly red or pink, indicating a ferruginous mineral; or white, indicating very pure quartz.)

Arkose quartzite or arkosite

Graywacke quartzite

Slaty quartzite

Schistose quartzite

Gneissic quartzite

} Indicating gradations

C. Special terms.

Taconite; the iron-bearing formation of the Mesabi Range. It is the protore and varies in both texture and minerals. The chief chemical constituents are iron oxide and silica, and the commonest textural feature is a cherty matrix with more or less well-defined granules about the size of a pin head (Fig. 190).

Jaspilite; banded chert and iron oxide of the other iron-bearing formations of the Lake Superior region.

Itabirite; quartzite with specular hematite.

Itacolumite; flexible sandstone.

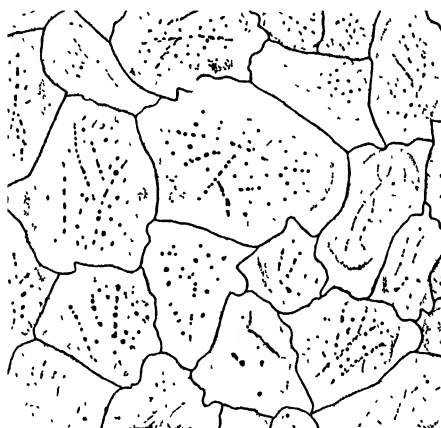


FIG. 231.—Sketch of thin section of quartzite in which the grains of quartz are enlarged by secondary growth after the sand was deposited. $\times 70$.

Notes on Quartzites.—The minerals of quartzites are comparable with those of sands (page 286), but some, such as the clay minerals, are almost wholly recrystallized to new minerals before the metamorphism reaches the stage of quartzite.

The typical quartzite mosaic texture somewhat resembles granitoid texture and commonly results from secondary growth rather than recrystallization in the body of the sand grains (Figs. 231 and 232). If the quartz has recrystallized under stress, a certain amount of elongation of grain and parallelism may be seen. We have no good textural term to distinguish the large grains residual from granulation from the crushed grains in the mosaic or sutured matrix. The luster of a quartzite is characteristically more vitreous than that of sediments.

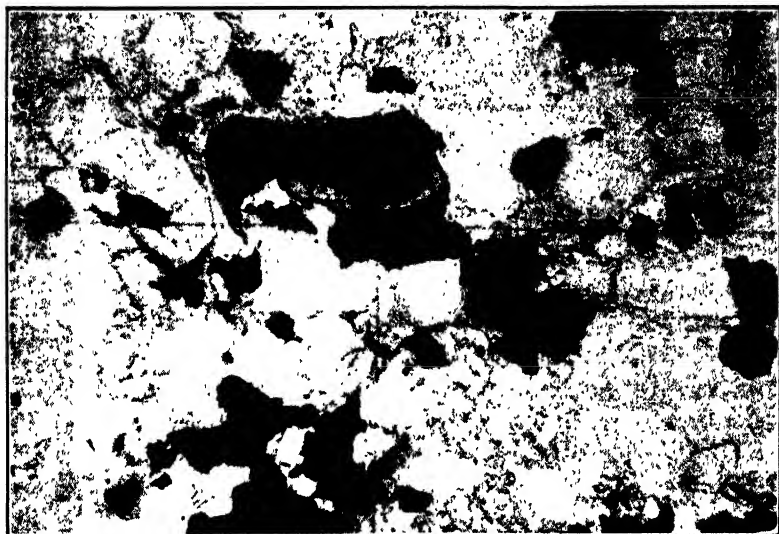


FIG. 232.—Photomicrograph of arkose quartzite, Mufulira Mines, Rhodesia. A tourmaline sand grain shows secondary growth. Plain light. $\times 100$.

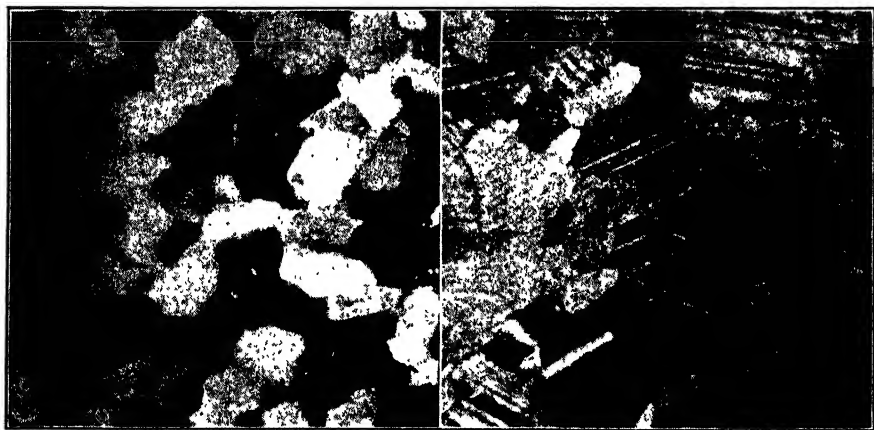


FIG. 233.—Photomicrograph of quartzite mosaic resulting from recrystallization without much deformation. New Ulm, Minnesota. Crossed nicols. $\times 10$.

FIG. 234.—The mosaic of carbonate grains in Carrara marble. Crossed nicols. $\times 100$.

METAMORPHIC CARBONATE ROCKS AND CARBONATE-SILICATE ROCKS

(Fossils usually destroyed)

*A. Textural and structural varieties.***Granoblastic marble.****Banded marble and brecciated marble.****Slaty and schistose marbles;** usually owe their structure to some mineral impurity.

The type of "marbled" structure is irregularly streaked with greenish serpentine, etc. The texture is granoblastic. "Crystalline limestone" suggests the same texture but is less definite, including rocks that developed the texture diagenetically rather than by metamorphism.

Certain silicate minerals as they increase in abundance develop characteristic textures. Serpentine may be in rounded aggregates or veined with cross fibers. Talc is commonly platy to fibrous. Garnets are more nearly euhedral and may form metacrysts.

B. Mineralogic varieties. (Prefix mineral name to "marble," or "rock," or name of rock.) (Prefix color terms.)**Marble and dolomite marble.****Serpentine marble, ophalcalcite and serpentine rock.****Garnet rock and garnet zones.****Talcosed marble and soapstone.****Amphibole marbles and amphibolites.***C. Special terms.*

Cipolin; muscovite marble.

Epidosite; mostly epidote.

Predazzite; calcite and periclase.

Minerals of Carbonate Rocks.—The metamorphism of calcite and dolomite recrystallizes these minerals (Fig. 234) and may decarbonate the magnesite to form periclase. If impurities are present, the minerals may become complex and numerous—serpentine, talc, garnet, epidote, chlorite, chondrodite, amphiboles, pyroxenes, micas, graphite.

Carbonate rocks do not have any Al_2SiO_5 , staurolite, cordierite or brittle mica; they have olivine, wollastonite and humite only if there is shortage of alumina (Figs. 242 and 245).

Magnesium carbonate is dissociated and reacts with silica at lower temperatures than calcite, so that silicates of magnesium are often found with a carbonate almost free from magnesia.

The garnets in such rocks are andradite and grossularite, not almandite or pyrope. If iron and magnesia are present they form pyroxene rather than garnet.

CONTACT ROCKS

Most of the rocks produced locally by the heat and emanations from an igneous magma can be classified under gneisses, schists, quartzites, and marbles; they have much the same character near the igneous rock as farther away. A few additional types occur within this zone, however,

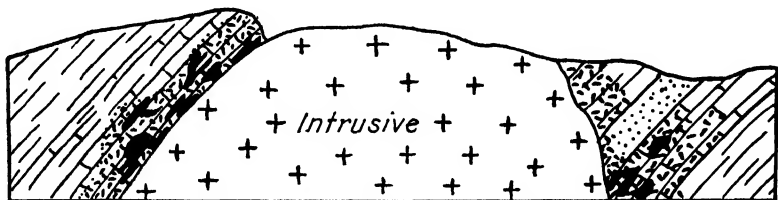


FIG. 235.—Hypothetical section of intrusive with contact metamorphism, and some ores (black) in the limestone. The metamorphic zone is much wider in the limestone than in the shale, and wider in both than in the quartzite.

and are not found more than a few thousand feet from such a contact (see Fig. 235).

Tactite,¹ a contact-mineralized rock with various minerals, formed from

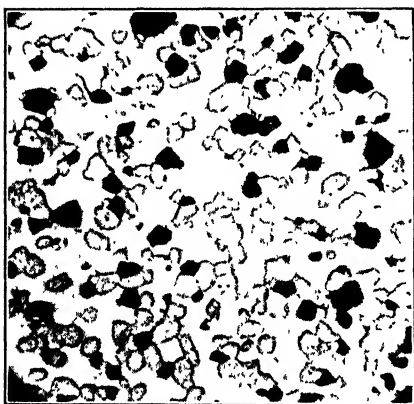


FIG. 236.—Photomicrograph of thin section of hornfels, Duluth, Minnesota. The minerals are plagioclase, augite and magnetite, recrystallized by the heat of a large gabbro magma. $\times 30$.

limestones or other soluble rocks by igneous emanations. Varieties are to be shown by mineral prefixes.

Skarn, a silicate complex associated with iron ores of contact origin; may be largely tactite.

Adinole, **spilosite**, and **desmosite** are albitized shales near the contacts of intrusives, usually the tops of diabase sills. Adinole is almost massive, spilosite has knots in a streaked hornfels and desmosite is much banded. If the newly formed feldspar is intergrown with wormy quartz, the mixture is symplektite.

Greisen and **luxullianite**. See hydrothermal rocks (page 372).

Hornfels, dense, somewhat cherty-looking shale, silicified by contact action.

Hornfels is also more broadly used for dense sugary-grained rocks, no matter what the original may have been, recrystallized by contact action about as thoroughly as a schist, but under such uniform pressures as not to show much schistosity even if it once had a schistose structure. The texture indicates rather a tendency for all the essential minerals to develop equidimensional or rounded forms, a granoblastic or granulitic texture (Fig. 236). Less commonly some

¹ HESS, FRANK, Tactite, Am. Jour. Sci., 4th ser., vol. 48, pp. 377–378, 1919.

rounded granules in a hornfels may be enclosed in a coarser mineral developed poikilitically¹ (Fig. 237).

Hornfels is a textural term then for rocks that have grains that are uniform and commonly not larger than 1 mm. It includes a host of mineralogic varieties, most of them named by prefixing to the term hornfels the minerals considered essential² (see page 392 and the diagrams). Even a granite or quartz porphyry may be recrystallized at high temperatures and develop a texture like hornfels.

Cornubianite is a tourmaline hornfels.

Hallefinta is much like a hornfels but some of it has phenocrysts, probably derived from a tuff.

Porcellanite is a very dense, light-colored hornfels.



FIG. 237.—Hand specimen of cordierite hornfels, Vredefort dome, South Africa. The large cordierite metacrysts weather out as knobs though filled with inclusions of other minerals. About one-half natural size.

Hornfels Minerals.—The expected minerals in shale hornfels are quartz, feldspars, cordierite, andalusite, sillimanite, mica, chlorite, garnet, staurolite, corundum, spinel, hornblende, epidote, rutile, anatase, ilmenite, graphite.³ Certain groups of minerals, however, are incompatible, as was true in igneous rocks (page 31). Quartz and corundum,

¹ Dana proposed the term metadiabase for a rock metamorphosed to a texture resembling diabase, but to most petrographers this term suggests a rock altered from an original diabase.

² Goldschmidt gives a series of ten in the range of altered sediments from clay to limestone; see "Die Kontaktmetamorphose in Kristianigebiet," 1911.

³ MacGregor, A. G., Clouded feldspars and thermal metamorphism, *Mineralogical Mag.*, vol. 22, pp. 524–538, 1931. Clouds of opaque dust are attributed to contact heating, not to deuteric or dynamic action.

for example, would react in a hornfels to form andalusite until one of the two was eliminated.

Rutile occurs especially in aluminous rocks.

In a "diabase hornfels" expect zoisite, epidote, vesuvianite, prehnite, garnet, albite, glaucophane, chlorite, rutile, kyanite.

In limy rocks there are also calcite, wollastonite, vesuvianite, diopside, grossularite.

In magnesian rocks expect olivine, spinel, and periclase. Limonite becomes specularite or magnetite. Organic matter grades to graphite. It should be noted that in mixed rocks magnesia enters into combination in silicates more actively than lime.

A great variety of igneous minerals and ores may be added as emanations from the magma.

HYDROTHERMAL ROCKS

Propylites are dark aphanitic igneous rocks, in which a green color has been developed by a characteristic alteration by hot water to chlorite, epidote, sericite, pyrite, quartz and calcite.

A further alteration yields more sericite. Even a mild alteration of orthoclase rocks gives sericite, but we have no special name for such rocks. A maximum intensity of hydrothermal action may locally silicify the rock.

Greisen is a granite in which such action has been so intense that nearly all has changed to quartz and muscovite.

Luxullianite is a similar aggregate of quartz and tourmaline.

WEATHERED ROCKS

Weathered rocks are described under the heading of sediments, pages 271, 272 and 297. They are chiefly soils, subsoils, and saprolites.

SUMMARIES OF METAMORPHIC FEATURES

Density and Composition.—The *density* of metamorphic rocks in the deep zone tends to become greater than that of the sedimentary equivalent, and in some greater than that of an igneous rock; as for example when a basalt (3.00) becomes an eclogite (3.50). On the other hand, rocks metamorphosed at small depths are heavier than sediments but lighter than the igneous rocks most closely related to them.

The *compositions* of metamorphic rocks are somewhat modified from those of the originals, but in general metamorphism (below the belt of weathering) comparatively few rocks are greatly modified, except as to water content. In contact action there may be large additions. No

TABLE XXI.—ANALYSES OF CERTAIN METAMORPHIC ROCKS

	1	2	3	4	5	6	7	8	9	10	11
SiO ₂	72.93	61.38	51.95	51.53	63.04	63.88	0.76	63.29	40.42	92.00	42.63
Al ₂ O ₃	12.67	15.24	12.58	10.61	16.45	17.70	0.42	1.24	1.86	4.21	1.53
Fe ₂ O ₃	None	3.31	0.90	2.98	1.32	3.02	0.16	2.75	1.80	31.41
FeO.....	2.08	3.43	8.77	9.93	4.89	1.80	4.68	4.27	0.30
MgO.....	0.62	2.61	8.90	8.74	5.04	3.72	1.01	27.13	35.95	None
CaO.....	1.91	4.79	7.00	9.97	3.17	2.72	54.67	Trace	0.66	0.04	23.37
Na ₂ O.....	3.19	3.29	2.79	2.62	2.62	1.78	0.16	0.16
K ₂ O.....	4.55	2.05	1.38	0.91	2.14	3.34	10.51	0.96
H ₂ O+.....	0.50	2.38	2.67	0.67	0.66	0.87	4.40	0.21	None
H ₂ O-.....	0.11	0.14	0.08	0.03	0.23	None	0.14	Trace
TiO ₂	0.54	0.50	1.03	0.60	0.54	0.52	Trace	0.21
P ₂ O ₅	0.11	0.70	0.24	Trace	0.58	1.44
CO ₂	0.38	0.02	1.02	0.57	43.49	1.24	0.43
Miscellaneous.....	0.21	0.26	0.43	0.71	0.48	0.45
Sp. gr.....	99.80	99.96	99.56	100.16	100.38	100.61	100.35	100.90	99.47	100.68	99.67
	2.665	2.956	2.735	2.680

1. Granite gneiss, Argentina. M. Dittrich, analyst.
2. Diorite gneiss, British Guiana. J. B. Harrison, analyst.
3. Chlorite schist, Lake Superior greenstone. S. Darling, analyst.
4. Hornblende schist, Lake Superior greenstone, near granite. F. Grout, analyst.
5. Biotite schist, Lake Superior, from slate. F. Grout and S. Darling, analysts.
6. Slate, Lake Superior, Knife Lake slate. D. Manuel, analyst.
7. Marble, Georgia. W. H. Emerson, analyst. Other marbles may be dolomitic.
8. Soapstone, steatite, South Africa. Van Riesen, analyst.
9. Serpentine, Massachusetts. George Steiger, analyst.
10. Quartzite, South Mountain, Pennsylvania. F. A. Genth, analyst.
11. Garnet rock, Clifton, Arizona. George Steiger, analyst.

extensive compilations of analyses of metamorphic rocks or averages of large collections are available, but the table here given is believed to be indicative of the general range of the rocks. The range in each class may be estimated by a study of the large collections of analyses of igneous and sedimentary rocks.

Economic Interest in Metamorphic Rocks.—The economic features of metamorphic rocks are less important than those of sediments and igneous rocks. As building materials quartzite, marble, serpentine and gneiss are harder and in several respects fully as good as the rocks before metamorphism; slates for roofing and other uses are of no value until considerably metamorphosed. Common quartzite is more difficult to quarry and trim into blocks than sandstone, and the cleavable schists are too easily cleaved for structural purposes.

Certain minerals such as garnet for an abrasive, and talc, asbestos, corundum (magnesite?), and graphite are derived chiefly from metamorphic rocks, and are products of metamorphic action.

Among the ores iron is mined largely from leached metamorphic rocks; but few ore deposits are much improved by dynamic metamorphism. Contact metamorphic ores are important in a few places but make a small proportion of metallic ores in general. Many veins of ore are associated with hydrothermally metamorphosed rocks, so that this type of metamorphism acquires a good deal of economic interest.

PART VII

THE PETROLOGY OF METAMORPHIC ROCKS

GENERAL FEATURES

Metamorphism is broadly defined as any change, physical or chemical, in a rock. As practically all rocks, however, have been modified to some slight degree since they formed, the term metamorphism is commonly



FIG. 238.—Photomicrograph of thin section of gabbro gneiss with albite twinning bands (which grow straight) now bent by deformation of the solid rock. The associated wavy extinction and crossed twinning are also interpreted as signs of deformation. $\times 100$.

reserved for those extensive changes involving new crystallization or new textures throughout the body of the rock—usually those changes below the belt of weathering involving higher temperatures than those normal at the surface. (See the further readings listed.)

It is known that metamorphism of sediments at depth may produce minerals and textures somewhat resembling those of igneous rocks, and there have been many suggestions that certain igneous rocks represent an extreme effect of the processes of metamorphism. It is believed, however, that metamorphism is seldom so extreme that the record of metamorphic action is lost, and that it is even more rare for a rock in the

earth's crust to be melted to form an igneous rock—the process Sederholm calls palingenesis or anatexis. Most gradations from metamorphic rocks to igneous rocks are a result of injection and assimilation rather than gradual fusions. A few rocks, however, strongly suggest partial fusion or hydrothermal fusion of some deeply buried originals (see pages 171 and 189).

EVIDENCES OF DEFORMATION

A high proportion of metamorphic rocks show structures in a large way indicating that the rocks have been deformed. Folds and faults and highly tilted structures appear in formations that seem to have originated only in relatively flat uniform sheets. In smaller specimens not only some small faults and folds but other structures are found in forms never seen as a result of original growth, Fig. 238. By experiment also it is found that deformation produces in some minerals optical “strain shadows” and in others a twinning; in some rocks a cleavage or schistosity and augen structure. Furthermore, after long study it has been learned that certain minerals are found chiefly in deformed rocks, and that the proportionate elongation of certain minerals is greater in deformed rocks than others. These minerals and forms thus become criteria of deformation and are listed below.

CRITERIA SUGGESTING DEFORMATION AT SHALLOW DEPTHS

Faulting and folding.

1. Slickensides.
2. Gouge; claylike, in many places containing fragments of the wall rock blackened and slickensided.
3. Drag folds and minute crenulations.

Brecciation and shattering.

1. Autoclastic or friction breccias. These are identified by (a) angularity of fragments, (b) homogeneity of fragments, (c) vein cement.
2. “Crackle breccias” or breccias in which the fragments are not rotated.
3. Veinlets, numerous, closely spaced, many in sets.
4. Offset of bedding laminations or other structures.
5. Fracture cleavage as expressed by rhombic cleavage blocks.

Cataclastic metamorphism. (Produced by directed pressure or stress without the effects of heat or recrystallization and hence usually at small depth.)

1. Strain shadows, especially in quartz, and banded; especially if very pronounced; biaxial quartz.
2. Irregular twinning effects: (a) Patchy twinning, (b) wedge-shaped or interrupted by cracks, (c) curved, (d) lamellar twinning in micas, (e) microcline structure developed in orthoclase.
3. Irregular birefringence.
4. Bent crystals.
5. Slicing, or breaking and slip of crystals along parallel planes.
6. Distortion of original forms and structures: (a) of pebbles, (b) of volcanic structures, amygdules, spherulites, ellipsoids; (c) of fossils, (d) of crystals, etc.
7. Pebbles with transverse quartz veins that do not penetrate the conglomerate matrix.
8. Pebbles with stylolitic contacts with one another and with shallow circular concavities at point of contact.
9. Fractured crystals: A fracture running across several crystals; a series of inclusions indicating a fracture filled up and healed; a series of fractures across the

minerals of a rock in one general direction (especially following several cleavage planes as it crosses a mineral with a zigzag course).

10. Schiller structure? in pyroxenes, feldspars and olivine.

Granulation, crushing, and shearing.

1. Mortar structure, or the granulation of the borders of crystals leaving the cores intact.

2. Augen, large crystals with flattening, elongation and peripheral granulation.

3. Flaser rocks, augen of rock rather than coarse crystals.

4. Pulverization, with formation of a structureless aggregate of fragmental material of various sizes.

5. Mylonite structure, or complete granulation with differential mass movement, giving well-marked lenticular or parallel bands of contrasting minerals.

6. Hartschiefer, ultramylonites partly recrystallized; banding of the most rigid parallelism and equality of thickness with a cherty or felsitic appearance.

7. Pseudotachylite, local fritting and partial fusion, forming isotropic black streaks.

8. Secondary pyrite in a limestone is more abundant near faults (see Somers, R. E., *Petrographic criteria of structure*, Econ. Geology, vol. 23, pp. 317-322, 1928).

CRITERIA SUGGESTING DYNAMIC METAMORPHISM (DEEP EFFECTS)

(Produced by direct pressure and heat, which cause recrystallization and the formation of new minerals and textures. Those marked by * are in part static and plutonic.)

Textures and structures.

1. *Shapes* of metamorphic minerals: Anhedral, lenticular, rounded, sinuous and sutured; frequently a dominance of fibrous and platy minerals; metacrysts* euhedral; *index of elongation* (biotite and chlorite 1.5 in granite, more than 2.0 in dynamic metamorphic rocks; hornblende 2.5 in igneous rocks and 5.0 in schists).

2. *Textures*:* (a) Confused aggregates of fibrous minerals (amphibolites), (b) sieve texture, (c) *helizitic* texture, or the inclusion of oriented older minerals in a metacryst, (d) *diablastic* texture, or interpenetration of minerals of rodlike form, (e) *porphyroblastic, metacrysts* (often surrounded by a bleached zone), (f) mafic minerals later than quartz and feldspar.

3. *Structures*: (a) *Schistose* and slaty, flow cleavage, silky luster, parallel arrangement of elongate and platy minerals, (b) *Gneissose*, banding and foliation, an alternation of schistose and granular bands which differ in color, mineral composition and texture, (c) *Granulose*, sutured grains (as in quartzite). Mosaics.

4. *Signs of recrystallization*: See criteria page 411.

5. *Crystals*: (a) Simple crystal forms generally related to cleavage planes, (b) crystalloblastic, holocrystalline, (never cellular or glassy); with all crystals growing at once, some grow more nearly euhedral. Order of tendency to form euhedral crystals, or crystalloblastic order (page 357). (c) Absence of skeleton crystals, (d) absence of zonal growth.

6. *Inclusions*: (a) All minerals *mutually inclusive* (except metacrysts), (b) inclusions generally more idioblastic than the host, (c) "snowball" garnets, with spirally arranged inclusions, (d) inclusions do not follow the strata of the crystal but are related to the pyramids or an older helizitic structure.

7. *Pygmatic folds*.

Minerals.

Minerals nearly always metamorphic (in part also deep veins) staurolite,* chloritoid,* glaucophane, lawsonite, crossite, uralite, andalusite, sillimanite, chlorite, talc,* serpentine, sericite, omphacite, anthophyllite, gedrite, actinolite, tremolite, zoisite, grunerite, margarite, birefringent garnet, cordierite, kyanite.*

Minerals usually metamorphic, especially when in large amounts: muscovite, garnet, epidote, graphite, specular hematite, corundum, spinel, beryl, titanite, apatite.

The anorthite end of the plagioclase series tends to form the outside of crystals in the metamorphic rocks. Amphiboles are more common than pyroxenes; kyanite* than sillimanite and andalusite; rutile* than brookite.

WHERE AND HOW ROCKS ARE DEFORMED

Rock flowage as distinct from fracture is a deformation under such conditions that no visible openings result, and the rocks retain their coherence and most of their strength. Even in rock flowage, however, crystal grains may be broken and sometimes crushed to powder, though under the prevailing pressures that powder is simultaneously welded into a coherent mass.

Anticipating for a moment the detailed statement of rock flowage (page 401), it may be noted that this process involves no fluidity in the

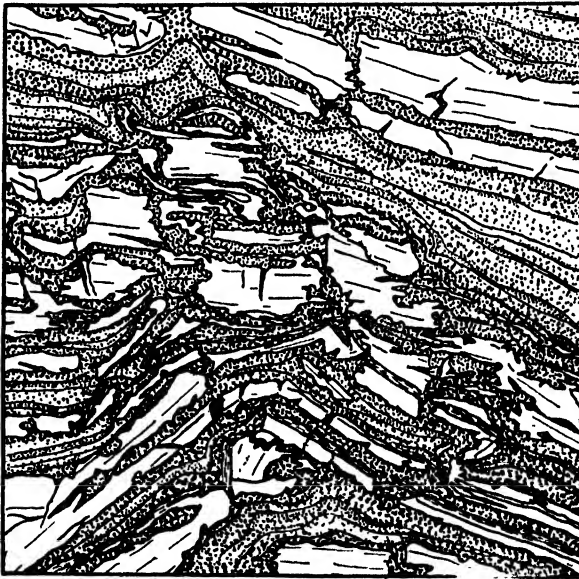


FIG. 239.—Jaspilite deformed by fracture and flow. (After Van Hise and Leith.) The jasper layers were more broken than bent, but the hematite flowed into the cracks and adjusted itself to the spaces left. About natural size.

mass as a whole. Slight pressures change the form or volume of the rock "within the elastic limit." Larger pressures deform the rock permanently (1) by a "granulation" or crushing of the crystals, and (2) by intermediate effects known as gliding and twinning. (3) Large stresses, and probably small stresses if prolonged, cause "recrystallization."

Weak rocks like clays and glacial ice yield to moderate stresses even at the surface of the ground and not only fracture but flow. And stronger rocks also yield to moderate stresses, if the stresses are prolonged as in the window caps of very old buildings.

Deeper in the earth some rocks under stress are fractured, others flow (Fig. 239), and there may be conditions or depths at which all rocks yield by flowage. The chief yielding of rocks that involves new crystallization

occurs (1) in the deeper parts of the earth where the results are forever lost to observation, (2) in the regions of mountain folds, and (3) especially in those mountains invaded by batholiths, so that the pressure is reinforced by heat and mineralizers. Rocks in such locations commonly develop a flow cleavage, although under the most intense conditions they may so far recrystallize as to become much like igneous rock (page 171 and Fig. 240).

The type of metamorphism that produces fracture cleavage probably involves very great thrusts but not such great confining pressures, depths or temperatures as those producing flow cleavage (see page 429).



FIG. 240.—Fan folds with cleavage following the bedding, Georgian Bay, Ontario. Characteristic of deformation under very intense heat and at great depths.

FACTORS OF METAMORPHISM

The factors involved in metamorphism are here grouped as forces, agents, the nature of the rock affected, and time. The forces are heat, uniform pressure, non-uniform pressure or stress, chemical affinity, and crystal forces. The agents are magmas, water solutions of great variety and gases, especially when hot or in solution. Some of these factors may be noted briefly, but others require more careful discussion.

Heat.—The sources of heat for metamorphism are about the same as those considered in connection with magmas (page 145), but the emphasis is commonly put in a different place. Given some internal earth heat, establishing a geothermal gradient, the heat that is a factor in metamor-

phism is believed to have been applied to rocks as they become depressed in geosynclines to depths where high temperatures are normal. Any heat in excess of the normal gradients is attributed to magma injection or very locally to friction.

Magmatic heats vary with the several magmas. Superheat, more than is needed for the maintenance of a liquid state, has rarely been noted except in basaltic magmas. These magmas may rise from considerable depths, bringing the temperatures that would be normal at those greater depths. On the contrary, other magmas are probably often derived from basaltic magma by the separation of crystals. If so they are probably saturated with some constituent during their whole evolution and are not to be considered as bearing much superheat unless locally reheated.

Temperatures of magmas range from 1300°C. in basalt lavas in craters down to 500°C. or less in pegmatite magmas. The temperatures of metamorphism range from these down to the temperatures of the rocks at the surface.

Rocks are poor conductors of heat (see page 151). If a magma at 1000°C. were injected against a limestone at 0°, the heat would be conducted into the wall so slowly that in 100 years the rock 100 meters from the contact would be heated only to 110°C. Heat is probably carried into such a rock, however, much more rapidly in many cases by the emanations of solutions from the magma. This action is convection, as distinguished from conduction, of heat. In regional metamorphism by the rise of heat in a sinking geosyncline, the conduction must be very slow, but the time available is no doubt very great. At depths down to 10 miles the general temperature is probably not more than 500°C., and any signs of higher temperatures suggest igneous intrusives. (See the criteria of the geologic thermometer for other evidences of high temperatures.)

The effects of heat in metamorphism are outlined on pages 394 to 396.

Pressures.—Pressures in metamorphism are largely the direct or indirect effects of gravity. The only other possible metamorphic pressure is the expansion of a mass of rock when heated, causing pressure outward in all directions; and conversely a loss of heat after contact heating may produce tension cracks and veins in the contact-metamorphic rocks.

The common notion of gravity pressure is that of a simple load of overlying rock. An increasing load, from the deposition of a geosyncline of sediments or from a series of lava flows, causes increased pressure and a certain amount of compression. If a rock is not supported on the sides with a confining pressure equal to the load, some of the load is resolved into lateral forces tending to spread the heavily loaded rock sideways. During isostatic adjustments these lateral pressures move large volumes of rock.

It is very uncertain, however, whether either loads or isostatic thrusts cause any very intense metamorphism in the crust. Under heavy loads a clay may become a slate with cleavage parallel to the bedding, but such

slates are rare compared with those formed by lateral thrust. The lateral thrusts due to loads are at a maximum at depths so far below the crust (page 160) that the rocks affected are never seen; they lie in the zone that has been named an asthenosphere. The magnitude of the stresses under differences of relief depends on the excess of load. The shallow stresses are slight, the deep ones large. It has been noted also that burial under heavy loads, equivalent to several miles of rock, does not always produce much metamorphism (page 189). In Pennsylvania the structures indicate that sediments have been folded at depths of several miles with insignificant metamorphism.¹ Similar structures are recorded in the Cordillera and in the Canadian shield. Taking all these matters into account, it seems likely that the major effects of pressure in metamorphic rocks are caused by some force other than the direct load. Load and lateral adjustment due to a shift of load can hardly push up the folded and overthrust mountains.

Major orogenic movements are commonly attributed to pressures that depend on some deeper changes in the earth, such as the shrinkage of a core of the earth by cooling, or by condensation of other sorts, or by a conduction of heat from one zone to another. If some zone or sphere in the earth is reduced in volume, the vast shell of overlying material, possibly of great strength, may develop enormous lateral thrusts in adjusting itself to the core.

These vast forces have to be considered in relation to the earth as a whole. In Archean time deformation and metamorphism seem to have been almost world wide, though possibly not everywhere at once. In later periods "bow areas" were folded as arcs around the more rigid "shield areas." Still more recently the main deformation has been largely restricted to the "hinge zone" (see pages 131, 172 and 248).

To explain the localization of the deformation we have only to find a cause for local weakness under the general thrusting forces. The deformed belt may have been weakened by (1) weak rocks deposited as a sediment in a geosyncline, or (2) a weak crust below the sediments (whether a cause or an effect of the growing geosyncline), or (3) the rise of batholiths thinning and softening the crust.

The several specific regions of characteristic deformation are notably coincident with regions of batholithic invasion, but the nature of the connection is not so obvious as the fact that the connection exists. The roofs and walls of many batholiths are intensely deformed. The thrusting effects of magmas appear to be prominent in the Appalachians, where batholiths are in general opposite to the greatest thrusts, metamorphism and deformation (page 172). This is characteristic of the mountain type of batholiths and may be a result of the general orogenic thrust acting on

¹ BARRELL, JOSEPH, Relation of subjacent igneous invasion to regional metamorphism, *Am. Jour. Sci.*, 5th ser., vol. 1, p. 18, 1921.

some distant portion of the batholith magma, which then transmits the pressure hydrostatically to all parts of its wall. A detailed study of the sequence of deformation and batholithic action shows that in some places deformation began shortly before granite intrusion and ended before the last granite was injected (pages 439 and 440). For the most part injection and mashing overlap and alternate.

Thus although there is a connection between batholiths and folds, there are some batholiths without much folding near by, and there are extensive regions of folding without known batholiths. The development of thrust is not so localized in the roofs of batholiths as the great heat and abundant emanations, which are the other major factors of metamorphism.

In summary, it seems that the pressures of most significance in metamorphism are those great lateral thrusts related to orogeny and probably related to some deep-seated internal adjustments in the earth. Locally, in the roof and walls of batholiths in orogenic belts, there are thrusts probably of the same origin transmitted by the magma of the batholith itself.

The effects of pressure on surface exposed, on heat, on texture, and on deformation are described on pages 390 and 398 to 400.

Chemical Affinity.—There is a general scarcity of laboratory data that have any direct application to the problem of how the minerals of schists are made by recrystallization under pressure. Few schists have been made artificially. A large number of specimens have been deformed so as to flow but this effect involves little recrystallization and the products are not very schistose.

Even the schists produced in nature are not very thoroughly known. Some of the minerals most characteristic of recrystallization to schists have rarely been separated for analysis or test. Some of the minerals of variable composition may be derived from a great variety of original rocks. On the other hand, the chemical affinity must vary with conditions, for several metamorphic reactions are known to be reversible. Most of the reactions may be classed as one of the following, or some combination of them: rearrangements, hydration, dehydration, oxidation, reduction, addition, and metasomatism.

Crystal Forces.—It is known that growing crystals may exert a mechanical pressure on the material surrounding them, but it is hardly likely that their growth is an important factor in metamorphic pressures. The metacrysts of many schists transgress the structure of the rocks in a way to show that the growing crystal did not deform its neighbors. It is concluded that when the schistosity bends out around a metacryst the schistosity developed later than the metacrysts (see Fig. 261).

Crystal forces are probably involved also in the elongated habit of certain mineral groups.¹ The uniformity of this ratio for each mineral

¹ LEITCH, C. K., Rock cleavage, U. S. Geol. Survey Bull. 239, pp. 28-40, 1905.

RATIO OF GREATEST TO MEAN DIMENSION

Micas.....	as 10	is to 1
Amphiboles.....	as 4	is to 1
Quartz and feldspar.....	as $1\frac{1}{2}$	is to 1

in rocks recrystallized under differential stress indicates that the flatness depends on the presence of stress rather than the extent of deformation. The flatness does not increase with increasing deformation. Crystal forces also determine the apparent "crystalloblastic series" (page 356) and probably modify all recrystallization, pages 406 to 411.

Water. *Quantity of Ground Water.*—Water is present in nearly all rocks, and the process of recrystallization, which is the essence of much metamorphism, is commonly attributed to its action. Water is not necessary but in geologic phenomena its influence is rarely excluded. It was once supposed that below the belt of weathering all rocks had their intergranular spaces filled with water, but this has been found to be untrue. The circulation of meteoric water is in many regions limited by some relatively impervious bed, below which the pores have almost no water at all.

The pore spaces in average sandstone are about 20 per cent, and in some shales they may be equally abundant but much smaller. Limestones have less porosity, around 5 or 10 per cent, and igneous rocks commonly have less than half a per cent of pore space (though flows may be highly vesicular). Below the superficial, discontinuous sedimentary rocks the water of meteoric origin is probably largely stagnant. So far as water in pore spaces is concerned, minerals are movable mostly in upper zones and relatively immovable below. Most deep mines are dusty in their lower levels. Great salt beds underlie some regions of abundant rainfall and are not noticeably leached by circulating water.

On the basis of such facts the estimates of the quantity of ground water have changed very greatly. Delesse in 1861 estimated that rocks held in pore spaces water equivalent to a layer 7,500 feet deep over the globe; Schlichter in 1902 estimated 3,000 feet; Van Hise in 1904, 226 feet; Fuller in 1906, with a careful review of earlier opinions, estimated 96 feet; and Kemp¹ thought it was even less. Small amounts of water, however, may have important effects.

Sources of Water Involved in Metamorphism.—1. It has just been noted that the quantity of meteoric water in circulation in rocks, though much less than formerly supposed, may yet be large enough to facilitate metamorphism.

2. Probably the estimates just mentioned include a certain amount of "connate" water—water that filled the pore spaces of sediments at the time of their deposition as sediments.

¹ KEMP, J. F., Present trend of investigation on underground waters, *Science*, n.s., vol. 28, p. 352, 1908.

3. The minerals of rocks contain various amounts of water as part of their essential composition. When some highly hydrous mineral, such as kaolin with 14 per cent of water, is subjected to metamorphic heat, water is liberated, and it is believed that this water may be very effective as an agent in recrystallizing other minerals. The average sedimentary rock has 4.28 per cent, the average igneous rock 1.15 per cent, and the average metamorphic rock possibly about 3 per cent of combined water. Evidently it is mostly from the sediments that the water of composition can be liberated. As some sediments have been deeply buried and not metamorphosed greatly either by water in the pore spaces or by water from the hydrous minerals, it may be questioned if such water is the most effective.

4. Magmatic emanations (described at length on pages 216 to 220) are largely water. This magmatic water is particularly active on account of its high temperature, associated acid gases and the high pressure under which it comes out into other rocks. The fact that some profoundly altered rocks have not been deeply buried or much deformed but are near igneous intrusions argues strongly that emanations are more effective than other waters, though of course, magmatic heat may greatly activate water of meteoric origin. The presence of mineralizers other than water in rocks over some wide regions is supporting evidence that contributions were received from magmas. Even such a common mineral as biotite in metamorphic rocks contains some fluorine of uncertain origin. The water in such minerals may, of course, be a mixture, for the emanations enter rocks that already contain water. Nevertheless, it seems clear that many metamorphic products must be attributed partly to magmatic emanations.

Magmas as intruded probably contain from 1 to 10 per cent of water held in solution by pressure. If the pressure is reduced, some of this water may boil out as water gas or separate as a solution above critical temperatures, carrying various substances dissolved in it, constituting an "advance guard" of igneous emanations. These may later condense and mix with other waters.

Quite in contrast with this type of separation of water from a magma, there is supposed to be a complete gradation from magma through pegmatite magma, possibly above critical conditions, to water solutions in a liquid state (page 213). It may be that some fairly concentrated solutions of silicates in water penetrate the walls and roof of a magma by virtue of the reduced viscosity of this watery fluid and its high vapor pressure. Finally, there may also be after the last stages of magmatic crystallization a watery fluid that affects the surrounding rocks as a hydrothermal solution.

In this sense, then, no sharp line can be drawn between magma and water as an agent of metamorphism. *Lil-par-lil* injections and pegmatites serve to illustrate the supposed gradation. Possibly some such injections were magmas with only a little water, but probably most were formed by hydrous magmas greatly reduced in viscosity, and still others may have been precipitated from hot-water solutions such as deposit veins.

Condition of Water.—The temperature of the water involved in metamorphism is nearly always assumed to be high, because cold solutions have generally less metamorphic action than hot ones. Certainly silicates dissolve in water so much more rapidly near 200°C. than at ordinary temperatures, that most petrologists are inclined to attribute recrystallization to the hot solutions. The effect of heat on the viscosity of water accounts for much of the pervasive character of metamorphism. If water at 0°C. has a viscosity of 1, water at 90°C. has a viscosity of 0.2, and at 300°C. a viscosity of only 0.05. The question whether the water is gas or liquid at high temperature is here left open. The physical chemists still disagree as to whether the aqueous emanations from magmas under high pressures and with high concentrations of mineral matter are gases or liquids (pages 214 to 220).

The activity of water in metamorphism is naturally much modified by materials in solution, such as the magmatic gases and minerals in the deeper hot waters, and atmospheric gases, organic compounds and some mineral constituents in surface waters. Sulphuric acid is found in some waters near the surface but is more active in weathering than in deeper metamorphism. Considerable emphasis should be given in metamorphism to the solutions of CO_2 , which produce reactions chemically as well as in dissolving minerals. The greater the pressure the more CO_2 can be dissolved and attack the rocks. The nature of some of the solutions is indicated by the chemical results on rocks, the contact emanations adding silica, iron and various common elements of magmas, and the hydrothermal solutions adding potash and sulphur, but leaching out soda and lime (see Fig. 250 and pages 217 to 222).

Circulation of Ground Water.—Hydrostatic head causes such circulation as artesian flow. Diffusion, capillarity, and osmotic pressure may cause further effects. The emanations from magmas, however, have a pressure behind them of a different sort. Magmatic pressure may be transmitted in a high degree to the emanations (see pages 213 to 215). If the emanations are gaseous, their change of state from liquid solution to separated gas involves an expansion that may maintain pressures if not produce them. These pressures are of a greater order of magnitude than the hydrostatic pressures on artesian waters.

In one sense the filter pressing of a residual magma or mother liquor may be compared with the motion of water in the pore spaces of sands. There is one difference, however. If the magmatic mother liquor is straining into a dike fissure, the dike may be kept open by the pressure of the emanation itself; the opening may even be enlarged by that pressure. On the contrary, a sandstone is likely to have its pores and fractures sealed by a cement deposited from the passing solutions.

Circulation in general depends largely on porosity, and pore space decreases with depth (see page 337).

Magma.—The differences in metamorphic effects produced by different magmas are partly due to differences in temperature and in pressure of intrusion and partly to the gases and solutions that emanate from the

magmas. It is believed that deforming pressures are more often related to granite than to gabbro magma, so that schist is more common near granites and non-schistose hornfels near gabbro. But there are some prominent granites near which no conspicuous stress effects appear. In general gabbro magmas are hot and give off a little Mg, Fe, Ti, Cl, etc., whereas granite magmas are not so hot but give off larger volumes of mineralizers—waters carrying silica, iron, sulphides, halogens, etc. Naturally also if the magma effects a metamorphism by injection *lit-par-lit* or by a more pervasive invasion, the resulting rock will depend on the nature of the magma itself. The structures resulting from very intense magmatic action are noted on page 171. Milder and more remote effects suggesting that magmas have invaded the formations include a mineralization, especially if a high-temperature mineralization with zoned arrangement of minerals; and any strong recrystallization that is not accompanied by strong deformation; and all the signs of exomorphic contact action (page 423).

When magmas evolve (pages 245 to 255) from basaltic to granitic compositions they change almost completely in the minerals that are stable in or saturate the magma. Hence in general the series of stages of magma show a sequence of effects on inclusions and wall rocks.¹ "Any magma will tend to *make inclusions over* into the phase or phases with which it is saturated, in so far as the composition of the inclusions will permit" (Fig. 262).

The relations of aplites and pegmatites to igneous rocks intruding limestone are also very significant. Pegmatite magmas are almost certainly more hydrous than aplite magmas and behave very differently as the dikes pass out from the igneous rock to the limestone. Aplites continue as dikes, more or less modified by the structure of the wall, but not reacting notably with the wall. Where a pegmatite dike passes out of a granite wall into a limestone it almost immediately disappears as a dike in a bunch or chimney of contact metasomatic silicates (Fig. 47).

Gases.—Gaseous emanations from lavas in volcanoes actively corrode and alter the rocks that they pass. Little work has been done to show the nature of the alteration. Possibly some of the deeper-seated pervasive contact action may also be accomplished by gases, but other effects result from hot liquids. Certain gases may be liberated by reactions in the progress of metamorphism, notably the CO₂ from carbonates and sulphur from sulphides.

Atmospheric gases aid in the weathering of rocks but probably act largely in water solutions.

Nature of the Rock.—The nature of the rock affects the process of metamorphism in many ways. Under a deforming pressure some

¹ BOWEN, N. L., The behavior of inclusions in igneous magmas, *Jour. Geology*, vol. 30, pp. 513-570, 1922; especially p. 538.

minerals are readily granulated, others slip along cleavages and still others readily recrystallize and are seldom found with any sign of having been crushed. For example, limestones may be granulated, but nearly all have later recrystallized so completely as to destroy the signs of granulation. Again, it is quite unlikely that metamorphism will produce a schist from a rock that has not a composition suitable for the production of an elongated mineral. Rocks consisting of quartz and calcite without platy minerals develop only a poor rock cleavage, but if a little of the constituents of mica or amphibole is present, the same degree of metamorphism will result in mineral orientation and rock cleavage.

The mineral nature of the rock also determines largely the minerals that can be produced and the effects of heat and water. A hydrous rock is less affected by waters either hot or cold than an anhydrous rock, whereas a granite would be less affected by dry heat than a hydrous rock like shale. Of all rocks limestones seem to be most affected by contact metamorphism. This is probably because heat dissociates the CaCO_3 , producing the highly active quicklime, and because the solutions that are involved in contact action probably carry the acid constituents of the magma, which are more likely to react with lime than with siliceous rocks like shale and sandstone. For a list of the products formed by the metamorphism of common rocks see pages 429 to 431.

Time.—The details of the influence of time upon metamorphism are not well known. As the average pre-Cambrian rocks are metamorphosed more than later rocks, there is no doubt that time plays an important part. It seems likely, however, that this simply means that the old rocks had more chances to be deformed and to be invaded by magma.

Most processes of metamorphism may probably be assumed to be so prolonged that equilibrium is attained. The compression and deformation related to deep burial are prolonged because it takes time to deposit such a load. The heat resulting from magmatic invasion is prolonged because rocks conduct the heat away only very slowly. More rapid deformation may occur in connection with mountain-making thrusts, and some heat effects may occur near small intrusives, but it is believed that relatively little metamorphic rock is formed by these brief forces.

Summary.—In summary it may be said that the tendency of opinion in recent years is to emphasize two major factors in metamorphism. In the first place there are deformation and folding by lateral thrusts in orogenic revolutions. In the second place, with some possible connection with the first, there are igneous invasions with their abnormally high temperatures and abundant active emanations. These two factors have different effects on rocks, and their various combinations with each other and with the minor factors make the results very complex.

STRUCTURE AND GROWTH OF CRYSTALS

Recent research in the structure of crystals indicates that they are built of units, probably atoms and ions of the elements, arranged in a regular way, forming what is known as a space lattice. In contrast to this regularity liquids and gases seem to have their particles in no regular arrangement. The process of crystallization, then, is a process of arrangement of the atoms and their simultaneous attachment to others similarly arranged, forming a more-rigid, stable unit—a solid crystal. If the tendency to orientation is only partly successful or is later disturbed, the partly oriented material may give parts of a crystal in a reversed position, forming a twinned crystal.

When one inspects a space-lattice model built to illustrate the structure of a crystal, it is easily seen that particles are arranged so as to be more closely packed in certain planes than in others. This closeness of packing is believed to be related to some such properties as the cleavage and the development of prominent crystal faces.

The arrangement of particles in a crystal space lattice varies with the size of the particles involved, or their "sphere of influence," and with their "deformability." It has been found that in a surprising number of cases it is possible to substitute, in the space lattice of a crystal, atoms of some other element analogous to one in the crystal. Two elements thus capable of replacing each other in a crystal are said to be isomorphous elements, and the mixed crystal is an isomorphous crystal. If isomorphous elements (and radicles) have similar volumes, the substitution of one for the other involves little distortion of the space lattice. (See the further readings outlined.)

At the contact of two crystals of different orientation it is believed that there may be a zone in which the atoms or ions are in a disorganized state or position—essentially amorphous.¹ This may be the zone in which atoms are transferred from one crystal to another during recrystallization.

The orientation of needle-like or platy crystals during recrystallization occurs largely by the growth of crystals, the growth being more rapid in favored directions. Favored directions are related to stresses (page 400), but the relation is complex. Stresses may be rotational or non-rotational, and the direction of elongation of the rock depends not only on the direction of the stress but on the amount of rotation. It ranges from 45° to 0° from the direction of maximum stress if there is shear, and is 90° from the direction of stress under non-rotational or direct pressure. The elongation of the crystals during recrystallization is parallel to the elongation produced in the rock,² not necessarily normal to stress.

¹ JEFFRIES and ARCHER, "The Science of Metals," McGraw-Hill Book Company, Inc., 1924.

² Leith and Mead, in "Metamorphic Geology," p. 173 *et seq.*, give a careful discussion of the evidences for this conclusion.

STRENGTH OF CRYSTALLINE ROCKS

The physical properties of a crystal are probably determined more largely by the crystal structure than by the chemical elements in it. The element carbon, for example, forms graphite in hexagonal crystals, which are opaque and have a density of 2.1 and a hardness of 1; the same element forms diamond in isometric crystals, which are transparent and have a density of 3.5 and a hardness of 10. Probably not only the strength of crystals but their cohesion and the consequent strength of rocks are dependent in some way on the crystal structure, but the relation is not well worked out. The texture of a rock naturally modifies the effect of crystal structure. Data on strengths of rocks are experimental.

TABLE XXII.—STRENGTHS OF SOME TYPICAL ROCKS
Under Compression in Laboratory Tests

Rock	Pounds per Square Inch
Granite.....	6,000 to 40,000
Basalt.....	6,000 to 23,000
Quartzite.....	About 27,000
Sandstone.....	2,000 to 15,000
Limestone.....	6,000 to 25,000
Shale.....	500 to 4,000
Marble.....	6,000 to 25,000
Marble under the confining pressure of a steel jacket...	Up to 70,000

It is noteworthy that sediments in general are softer and weaker than the metamorphic equivalents, but the difference between granite and gneiss is slight. On the average, then, metamorphism tends to make rocks hard and strong.

Strength and Confining Pressure.—At depths in the earth where rocks are deformed by flowage the confining pressures greatly increase the stress required to break down the space lattice.¹ Probably at pressures equivalent to a depth of 20 miles the strength of most rocks is four or five times as great as at the surface. Other factors may modify this, however, as shown in the following paragraphs.

Strength and Time.—Pressure of moderate intensity deforms many rocks within the elastic limit, or so that when the pressure is released the original form and volume are restored. If such moderate stress, however, is maintained for a very long time, the rock may acquire a permanent "set," and a continuance of the pressure produces further deformation. Old tombstones and the window caps of old buildings are found to be bent, and as they have not been weakened by heat or rapid solution the effect is to be attributed to the long continuance of stress. Clearly many strong rocks yield under small stresses if those stresses are long continued (see page 161).

¹ ADAMS, F. D., and J. BANCROFT, Internal friction in rocks, Jour. Geology, vol. 25, pp. 597-637, 1917.

Strength and Temperature.—The effects of heat are clearly evident to the field geologist as he notes the increasing deformation of rocks near igneous contacts. The result is of course confused by the possibility that solutions as well as heat are involved in the contact effect, but even in deformation by recrystallization heat must play a considerable part (see page 407). If heat thus makes rocks yield to a stress more rapidly or makes them yield to a smaller stress, it produces the effect of weakness.

There are few experimental data to show the direct effect of temperatures near fusion on the strengths of rocks. Adams¹ found that a limestone enclosed in a steel jacket did not flow into a cavity under a pressure of 96,000 pounds per square inch at room temperature for two and a half months; but the same rock showed some movement into the cavity in 70 seconds, if that pressure was applied when the temperature of the rock was 450°C., though this is far below the temperature of melting or dissociation. It is known also that metals weaken near the melting temperature,² and many refractory clays³ and slags soften and yield to very moderate stresses when heated at medium temperatures. Experiments with ice show that it flows much more rapidly near the melting point than at lower temperatures.⁴ In a general way, then, it seems to be safe to assume that rocks yield rapidly at temperatures somewhat below the beginning of melting. It is much less safe to assume that they yield rapidly at half that temperature. Data on actual rocks are much needed.

Strength and Solubility.—In the presence of water, soluble rocks may yield more readily under stress than less soluble ones and more readily than dry rocks, but further data are needed. Marble has its strength increased by deformation in the presence of water.⁵

MODES OF CRYSTAL YIELDING

Aside from temporary deformations within the elastic limit, rocks yield by three distinguishable processes. Grubenmann and Niggli suggest for these the terms elastic, plastic, and blastic.

Clastic deformation is practically the breaking up of a rock or crystal into smaller units. In each fragment the crystal structure may be preserved, but the orientation of the fragments becomes confused. A single crystal may be so crushed as to yield a million fragments. If this crushing occurs in the deeper zones of the crust, the movement of the

¹ ADAMS, F. D., Depth of the zone of flow, Jour. Geology, vol. 20, p. 111, 1912. ADAMS and NICHOLSON, Experimental investigation into the flow of marble, Roy. Soc. Philos. Trans., London ser. A, vol. 195, pp. 363-401, 1901.

² PRIESTER and HARDER, Effect of temperature on the properties of steel, Am. Soc. Steel Testing, Trans., p. 439, 1927. A steel melting above 900°C. was only 13 per cent as strong at 550°C. as at 68°C.

³ BLEININGER and BROWN, Testing of clay refractories, U. S. Bur. Standards Tech. Paper 7, 1912.

⁴ DELURY, J. S., Experiments with model glaciers, Jour. Geology, vol. 34, pp. 266-274, 1926.

⁵ ADAMS and NICHOLSON, *op. cit.*

fragments on each other may be so restricted by confining pressure that no visible openings and faults result; near the surface visible openings and faults are common. Some orientation of minerals is accomplished during crushing, producing *augen* and slightly rotating some platy grains.

Plastic deformation involves some modification of the space lattice of the crystal. Flexible crystals may be bent beyond the elastic limit, and many crystals may be deformed by twinning.

A similar deformation may occur in crystals by "*gliding*" along prominent crystal directions (Fig. 241). This, however, hardly affects the space lattice. Glacial ice may flow in large part by such gliding in crystals, (see page 161 also).

Blastic deformation involves recrystallization, and the space lattice may be more or less completely destroyed or replaced, depending on the nature of the recrystallization. Some rocks recrystallize without much mineral change and may have, as nuclei of the new rock crystals, some of the original crystals unmodified. Other rocks by recrystallization entirely change the nature of the compounds present, and in the altered rock no original crystal remains.

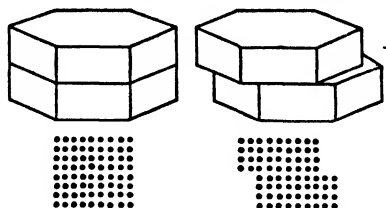


FIG. 241.—Simplified diagram of the manner in which a crystal may be deformed without much change in the space lattice. Hexagonal ice crystals may be deformed by gliding along the basal plane, without destroying the physical continuity or breaking the ice. (After Pirsson.)

EQUILIBRIUM

The Kinetic theory states that the condition of a system in equilibrium is not static or inert, but that action in one direction is equal to action in the opposite direction. If a crystal is in equilibrium with a liquid, it grows as fast as it dissolves. If certain reagents in solution are in equilibrium, the reactions in one direction are exactly as fast as those in the opposite direction. This principle is of the greatest importance in metamorphism.

In the first place, it explains much of the texture of the rocks produced. Every student of chemistry who has boiled a precipitate to make it coarser grained and more cleanly filterable must appreciate that the result is due to the solution and recrystallization of the precipitate. Solution occurs at all contacts of the crystals and liquids, and the small crystals have more surface per gram than large crystals; but crystal growth is more rapid on the large crystals than on the small ones. Thus with no appreciable change in its total amount the precipitate becomes coarser grained; the smaller crystals disappear and the large ones grow larger. Heat accelerates the process. In metamorphism, water, which occurs in practically all rocks, has apparently had a similar effect in

increasing the size of grain. Heat and long periods of time are favorable.

The universality of the effect of recrystallization is probably also related to the fact that no large quantity of water is needed. A little liquid in the space between minerals may again and again dissolve part of a grain and precipitate it elsewhere.

Attempts to interpret metamorphic equilibria on the basis of curves of melting and crystallization have very little value. Many of the minerals and textures of metamorphic rocks are not such as form from magmas. In one or two systems, however, the curves have some indirect applica-

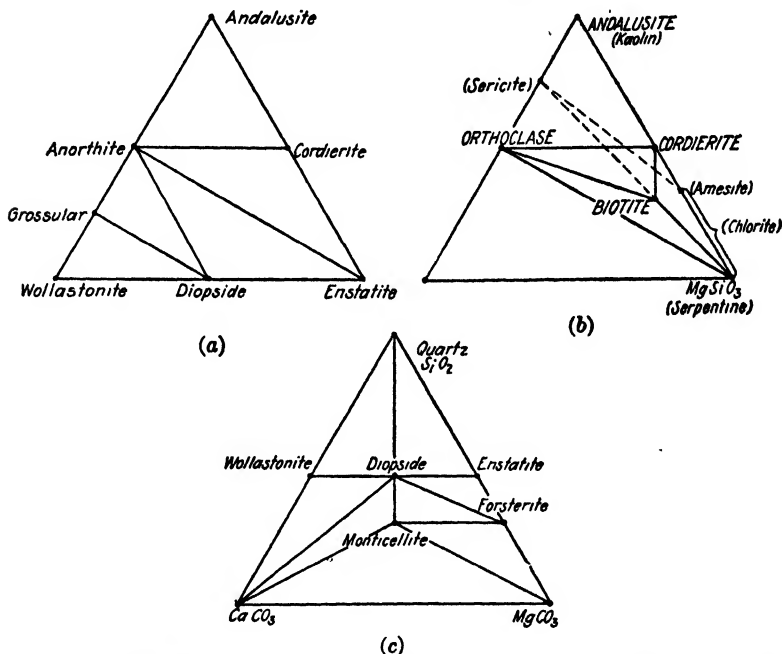


FIG. 242.—Diagrams of the probable equilibria of some ternary systems in metamorphism. (a) The shale-limestone hornfels. The system wollastonite-andalusite-enstatite. (After Tilley, *Geol. Mag.*, vol. 60, p. 411, 1923.) (b) The non-calcic shale hornfels. (After Tilley, *op. cit.*, p. 413.) (c) The metamorphic system $\text{CaO} \cdot \text{CO}_2$ - $\text{MgO} \cdot \text{CO}_2$ - SiO_2 in the kata-zone. (After Tilley, *op. cit.*, p. 416.)

tion to metamorphism, showing that the products of metamorphism could not have formed at such high temperatures as the melts.

Experiments on the crystallization of silicates from water solution or hydrous melts have not been carried far enough to be of much value as applied to metamorphism. In the absence of extended experiments the fields of stability of certain minerals in chemical systems with varying concentrations can be inferred from a study of the rocks themselves, the products of Nature's experiments, though the conditions of these experi-

ments are not accurately known as to either water content or temperature. Suggestions of the probable equilibria are shown in Figs. 242, 243, and 245.

The principles involved in these equilibria are known from experimental work on other materials, and several of the principles are significant in the discussion of particular metamorphic reactions. A few of the equilibria in metamorphism may involve gas phases, like the carbonate-silicate equilibrium, and others a system with two solids and no fluids. It is probable, however, that a high per cent of the new crystallization, which is the essence of metamorphism, involves a system of solid and liquid. The liquid is in some cases very hot but probably not hot enough

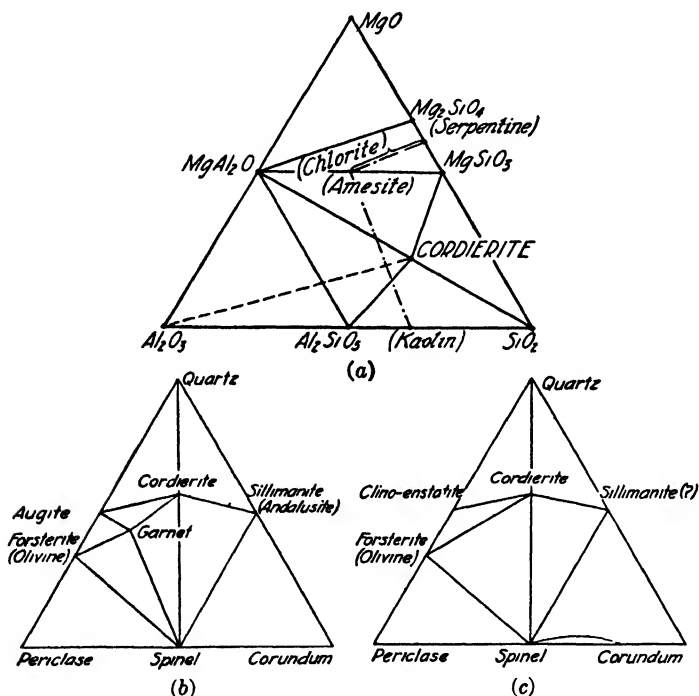


FIG. 243.—Diagrams of the probable equilibria in the ternary system $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$. (a) Hornfelses (probably in the meso-zone). (After Tilley, *Geology Mag.*, vol. 60, p. 415.) (b) Hornfelses in the kata-zone. (After Grubenmann and Niggli.) (c) In melts. (After Rankin and Merwin, *Amer. Jour. Sci.*, vol. 45, p. 301, 1918.)

to fuse the minerals. Practically all minerals are to some extent soluble in water and some water must be expected in nearly all geologic situations. The reactions of solutions on each other and on minerals are the mechanism of mineralogic metamorphism. The general principles applicable to these solutions are as follows:¹

¹ Mostly from JOHNSTON and NIGGLI, The general principles underlying metamorphic processes, *Jour. Geology*, vol. 21, pp. 481, 588, 1913.

Equilibrium and Reaction.—A system in equilibrium is completely defined by a statement of *temperature, pressure and concentration*. Reaction occurs when the equilibrium is disturbed by a change in one of these factors and may be reversed by a reverse change.

Incomplete Equilibrium.—Some minerals react so slowly to a disturbed equilibrium that the system gives a false impression of stability. It is believed, however, that most metamorphism, occurring at high temperatures and pressures and in the presence of hot water, attains at least approximate equilibrium in the long periods of geologic time.

Some metamorphism may be accomplished by passing solutions that do not constitute a "closed system" such as are dealt with in equilibria. For example, there are the solutions leaching silica from an impure quartzite, leaving sericite schist (pages 414 and 415), those adding silica and *lit-par-lit* and pervasive emanations; and those causing metasomatism. Any reaction in which some elements are wholly replaced indicates passing solutions rather than equilibrium. Absence of equilibrium is indicated also by the alterations that are controlled by volume relations, or by gel replacements (page 417).

Some systems of liquid-solid may on cooling develop a supersaturated solution for a time. In such systems the precipitation of a particular mineral is facilitated, and often the place of its growth is determined, by the presence in some part of the rock of a crystal of that mineral. It is common to find that quartz grows on quartz and calcite grows on calcite in the walls of a single veinlet.

Finally, it may be noted that the minerals, found in apparent equilibrium with each other in metamorphic rocks now exposed at the surface, are not in equilibrium with their surface surroundings. They formed at depth, and although they have long existed in shallow situations, the temperatures were not high enough to hasten reaction, and the time of exposure to weather before erosion is short except in a few regions of low relief. It is probable that the metamorphic rocks now exposed at the surface roughly indicate the equilibria during the maximum heat and pressure that they have experienced; though during erosion the effects of moderate conditions may modify those of the maximum intensity. It is suggested that some metacrysts that formed late in schists and gneisses may have grown during this transition from depth to the surface; and near the surface even the metacrysts may be altered.



HEAT AND ITS REACTION EFFECTS

Heat accelerates recrystallization whether water is present or not, but recrystallization from water solution is geologically most probable. Hot liquids dissolve substances more rapidly and usually dissolve larger amounts and being less viscous permit more rapid diffusion of dissolved salts than cold liquids. Cooling generally favors crystallization. Certain

complexities in the heat effects are noteworthy. For example, consider the decomposition of silicates by water solutions—one of the most common metamorphic reactions. At temperatures from 0° to 100°C. silica

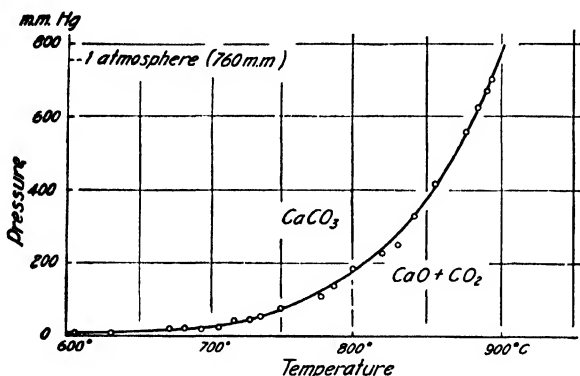


FIG. 244.—Curve of temperature and pressure of dissociation of CaCO_3 . At a little higher pressure the curve reaches 970°C. and there is a transition to $\alpha\text{-CaCO}_3$, which melts at 1290°C. (After Johnston, *Jour. Am. Chem. Soc.*, vol. 32, p. 942).

dissolves in ordinary water solutions much more easily and in larger amounts from silicates than it does from quartz; but at temperatures from 200° to 300°C. much more silica is dissolved from quartz than from common silicates. Furthermore, although alkali bicarbonate

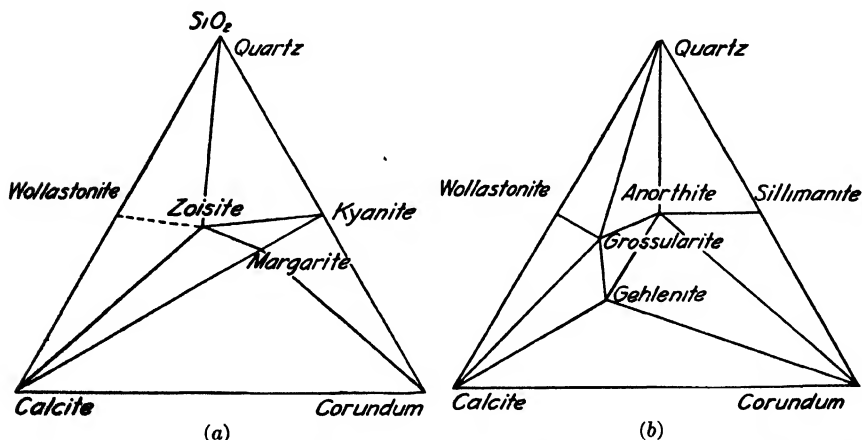


FIG. 245.—The metamorphic system $\text{CaO} \cdot \text{CO}_2 \cdot \text{Al}_2\text{O}_3 \cdot \text{SiO}_2$. (After Grubenmann and Niggli.) (a) In the meso-zone. (b) In the kata-zone.

solutions dissolve more quartz than would dissolve in pure water at ordinary temperature, the situation is reversed at high temperatures and distilled water dissolves more quartz than the bicarbonate solutions.¹

¹ GRUNER, J. W., Hydrothermal oxidation and leaching experiments, *Econ. Geology*, vol. 25, pp. 702-703, 1930.

Mineralogically a reversible reaction may be illustrated by:



if just the right quantity of each mineral be taken. If anorthite is in excess on the left, some will be found associated with hornblende in the product; but if olivine is in excess, olivine will be present with the hornblende. High temperatures move the reaction to the left and lower temperatures to the right. But probably water is also involved in the formation of hornblende, and some chance for water to escape is needed, if the hornblende is to break up into anorthite and olivine.

Another important group of reversible reactions are those of dissociation of carbonates and dehydration of hydrous minerals. Although heat drives out water and carbon dioxide, the reactions are opposed by pressure, so that hydration and carbonation may occur at great depths in spite of considerable heat (Fig. 244). Results in a particular case are hard to predict.

The differences in the products of reaction at different temperatures are also suggested by Figs. 243 and 245.

CONCENTRATION REACTIONS

The effects of concentration in most metamorphic reactions are not easily tested experimentally on account of the high temperature and pressure required. The general laws of such effects are difficult to apply to minerals of slight solubility and to silicates that readily break up into colloidal solutions.

The law of mass action is briefly that the speed of chemical action is proportional to the active masses of the reacting substances present at the time. Certain metamorphic reactions may serve to show the complexity of the applications of the law.

The reaction $\text{CaCO}_3 \rightleftharpoons \text{CaO} + \text{CO}_2$ depends not only on temperature and pressure but is immensely modified by silica.¹ In water solution at 260°C. silica displaces CO_2 from lime. The carbonate breaks up at the low temperature in spite of high pressure. In dry heat the series of products formed by CaCO_3 in the presence of SiO_2 is not well studied. There is some spurrite, $\text{CaCO}_3 \cdot 2\text{Ca}_2\text{SiO}_4$, and possibly some wollastonite.

In contrast with this suggested effect of dry heat it is found that in general the thermal effects are to be minimized (page 422). Nearly all the silicates formed in a sandy limestone are a result of reaction of the calcite with emanations, not with the sand.

Considerable importance should be found in hydrolysis, also, a reaction based on dilution rather than concentration, or more technically on the concentration of the solvent water in the system. Hydrolysis

¹ BARRELL, JOSEPH, Relations of subagent igneous invasion to regional metamorphism, Am. Jour. Sci., 5th ser., vol. 1, pp. 181-183, 1921.

may be involved in the precipitation and alteration of some important sulphide ores.

Several petrographers have summarized the probable effects of various oxides in a series of equilibrium diagrams (see Fig. 243 and the further readings listed). The series clay to limestone includes about 10 mineralogic intermediate rocks (page 371), and other mixtures are equally interesting. Some are complicated by emanations, however.

A well known qualitative rule is that *many dissolved salts modify the solubility of other salts*.

The most simple illustration of the effects of a dissolved salt is given by the presence of an ion common to both the solution and the salt. A water solution of a sulphate will dissolve less of another sulphate than will dissolve in pure water. To make the case general, in the reaction

$AB \rightleftharpoons A + B$, the law of mass action gives $\frac{[A] \times [B]}{[AB]} = K$ (a constant),

in which the symbols indicate the molecular concentrations of the substances reacting. Any additions of A or B to the system in equilibrium would increase AB to maintain the constant. The amount of AB in solution is limited by its solubility, and the maximum value of $A \times B$, which cannot be exceeded without precipitation of AB , is referred to as the *solubility product*.

The effects of substances in water acting on a series of silicates and quartz cannot be simply stated, but the field evidences are that the rule applies. A solution in contact with a rock may dissolve or attack some minerals and at the same time deposit others or may dissolve a mineral in one place and deposit it in another, depending on the salts supplied to the solution.

The law of mass action is not the sole control of metamorphic reactions, for there seem to be many cases in which the volume of the original mineral determines the amount of new mineral precipitated. It is noteworthy, also, that the precipitation of colloidal solutions is influenced by other factors besides molecular concentration.

PRESSURE REACTIONS

*In general, pressure increases solubility and a release of pressure causes precipitation from water solutions.*¹ The physicochemical law is that when pressure is increased, reaction tends to produce products of less volume. In attempting to apply this law to metamorphism we are met with the uncertainty about the "closed system." Some of the products, like CO_2 or water, may escape from the system and largely modify the reaction. Nevertheless the rule is of value.

Again, when a compound exists in two or more forms with different densities, *the higher pressures tend to produce the more dense forms*. The

¹ Note that this in sharp contrast with the effect of pressure on fusion. Pressure usually tends to form crystals from a melt.

case is probably illustrated by the formation of sillimanite at greater pressures than those producing andalusite, but most changes to more dense minerals involve also some dehydration at depth, and there may be certain exceptions to the rule.

Two cases must be distinguished in the study of pressure effects on equilibrium, the case of uniform pressure and that of non-uniform pressure.

✓ **Load Effects.**—Uniform pressures act equally in all directions on a rock, like the hydrostatic pressure in a liquid. Approximately uniform pressures are exerted on rocks when simply buried under a load of other rock. This pressure is believed to accelerate slightly the processes of solution and thus promote some reaction and recrystallization. A sort of static metamorphism results with little deformation. Possibly clay becomes argillite, and many schists develop metacrysts.

CRITERIA SUGGESTING STATIC METAMORPHISM

Definition (Judd, 1889).—A variety of regional metamorphism brought about by the action of heat and solvents at high pressures, the latter being due to a superincumbent load and not induced by orogenic deformation.

1. *Minerals.*—Kyanite, andalusite, cordierite, staurolite, garnet, biotite, chloritoid.

The above minerals developed as metacrysts, which

- a. Transect foliation and show no orientation.
- b. May contain helizitic inclusions, especially quartz.
- c. May be surrounded by clear aureoles.
- d. May show zonal structure.
- e. Will show a later deformation especially well.

Metacrysts of mica, chlorite, and chloritoid are especially good; *i.e.*, they are minerals that have a good cleavage and elongation.

2. *Other minerals suggestive of static metamorphism.*

Sillimanite, sericite, hornblende, omphacite, hypersthene, olivine, spinel, rutile, ilmenite, chlorite, albite, potash feldspars, anorthite.

3. *Rocks formed.*

Mica schist, garnet mica schist, staurolite schist, hornblende schist, mica and hornblende gneiss, coarse gneisses with biotite, sillimanite, omphacite or cordierite, granulites, eclogites, granulitic gneisses, jadeite, leptyte, crystalline limestone, quartzite.

4. Any signs of lithification or cementation.

5. Interference and interlocking of grains, coarseness of grains or increased size of grain.

6. Optical anomalies, secondary twinning, etc., *e.g.*, anomalous orthoclase and microcline.

7. Schillerization.

8. Parallelism between planes of original stratification and schistosity planes especially if the rocks are flat.

9. Recrystallization previous to folding.

10. Proof that a recrystallized rock has been deeply buried.

11. Intensity of metamorphism decreases upward.

12. No relation between the degree of metamorphism or crystallinity and the amount of crustal deformation; *i.e.*, vertical schists are the same as flat ones.

✓ **Stress Effects.**—Non-uniform pressure, differential pressure or stress, gives a series of effects notably different from the effects of uniform pressure. If the stress is great and rapidly applied, the crystals are broken down to thousands of fragments. This breaking into fragments may speed up reaction and the attainment of equilibrium, partly by

adding heat of friction and partly by exposing more surface to solution and reaction. The more general effect of moderate stress or of long continued stress is the effect on solubility.

In so far as any portion of a crystal is strained to a greater degree than another portion, the part under stress is more rapidly affected by chemical action. To emphasize the contrast between this strain and simple pressure, note that hydrostatic pressure probably raises the melting point of all rocks, but non-uniform pressure invariably lowers the melting point. The effects of stress on solution are similar to those on melting.

Consider a solid-liquid system like quartz sand and water, in which the pressure on the sand is the weight of overlying rock and the pressure on the water is that of the column of water above it. This is a difference in pressure in different parts of the system, and the sand, being under the greater pressure, will tend to dissolve, relieving some of the pressure as it goes into solution in the water at lower pressure. This is the application of the general rule that pressure increases solubility.

But in this same system, in which sand grains are strained by the weight of rock above, the strain in any one grain is greatest near the point of contact with adjacent grains. Those portions of grains that are most strained dissolve more rapidly than others; and the solutions, with relatively slight migration or diffusion, deposit the dissolved quartz on less-strained parts of the grains (Fig. 246 and page 408). Quartzite is thus formed from sandstone. This is the application of the rule given in the preceding paragraph. It is the type of recrystallization under pressure, which dominates theories of metamorphism. The fact of recrystallization is proved in many rocks by the new minerals formed. Examples are found in practically every schist and gneiss.

At temperatures near the melting points of minerals differential pressure may cause local fusion of grains, somewhat similar to the local solution here described. The process has been tested only for one rock, ice, which has an exceptionally low melting point. If pressure is much greater in one direction than another at temperatures near 0°C., some parts of the ice crystals melt and move around to places of less pressure, there to recrystallize. The whole mass is deformed or flows gradually during this recrystallization. Probably few metamorphic rocks have been heated so near the melting points of their minerals as to flow by such a method, except very locally near igneous contacts. Solution effects are more important than fusion in rock flowage.

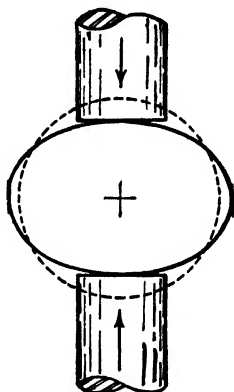


FIG. 246.—Diagram of the solution of material from a sand grain where most strained by a load, and its deposition on the less strained parts of the same grain.

Although few rocks have been tested in the laboratory by such deforming pressures in the presence of solvents or at temperatures near the melting point, enough materials have been tested to show the relative importance of uniform and non-uniform pressure. The effect of non-uniform pressure in ice is twelve times as great and in lead thirty times as great as the effect of uniform pressure. (It is a safe inference that much more metamorphic recrystallization is due to non-uniform than to uniform pressure.) This is the key, the *crux*, the *climax* of the whole study of the principles underlying metamorphic processes. Stress has been important in a large percentage of the metamorphism known. (See further readings listed.)

Textural Effects of Stress.—The textural effects of stress are, first, the reduction in size of grain during granulation; second, an increase in size of grain during recrystallization; and, third, an elongation of grains parallel to some plane of easy growth, a schistosity (page 382). It may be added that a rock once granulated by stress is probably more easily recrystallized by heat than one undeformed, and that this recrystallization does not require solution or fusion.¹

As far as the discussion has now gone the growth of crystals in the plane of the schistosity should occur without regard to the original position of the crystals. It is easily shown, however, by a study of the resulting rocks that some crystal forces are also involved in the result. Most of the elongated grains in a schist are optically and crystallographically as well as dimensionally oriented. It seems likely that such minerals as biotite in a schist are parallel because those plates in a position parallel to the schistosity found it easier to grow. The physical chemistry of this selective growth is not so well understood as the fact that schists almost always grow that way. Grains in other positions grow less rapidly, soon become subordinate and in that situation disappear as small grains usually do in recrystallization, serving to feed the growth of larger grains. The only other explanation that seems possible for the crystallographic parallelism is that grains may be extensively rotated under stress. Such rotation is hard to explain and is rendered improbable by the study of granulated rocks. Some of them show extreme granulation and a little rotation of plates and slices, but the textures do not closely resemble the schistosity that results from recrystallization. For the relation of schistosity to stresses see page 429.

PROCESSES OF METAMORPHISM

The metamorphic processes most often noted are *rock flowage*, *recrystallization*, and *granulation*. To a minor degree several sorts of changes of composition are also involved in metamorphism.

¹ JEFFRIES and ARCHER, *op. cit.*, p. 117.

Some of these processes are so different from others that several kinds of metamorphism have been distinguished. The chief of these are *anamorphism*, indicating the constructive or integrating processes whether by rock flowage, cementation, or replacement, regardless of depth; contrasted with *kalamorphism*, the destructive processes, chiefly weathering, which is usually excluded from metamorphism. Some distinction may also be made as between dynamic metamorphism, in which the mass is notably deformed, and static metamorphism, in which the dynamic effects are slight; also between thermal-, hydro- and chemical metamorphism and combinations. Most of these are not especially useful terms. The kinds most often mentioned are, probably, regional, contact and hydrothermal, and for a discussion of the factors involved these will suffice (see pages 421 to 427).

ROCK FLOWAGE

A rock is said to have flowed when deformed without conspicuous fracture, remaining at the end of deformation an integral body. This reference to "conspicuous fractures" means that the minute fracturing involved in granulation may result in true rock flowage. Usually, however, flowage is dominated or at least accompanied by recrystallization. Regional metamorphism is almost if not quite equivalent to rock flowage. Dynamic metamorphism is rock flowage.

Rock flow should, of course, be sharply distinguished from igneous magmatic flow, though in a few cases the results may be so similar as to leave the history of the rock in doubt. Igneous flow may continue until the magma is largely crystalline and the crystals well oriented, giving trachitic and trachitoid textures. If deformation and true rock flowage occur before the newly solidified rock is cold, Weinschenk believes that the metamorphic rock is distinguishable from a rock metamorphosed at a much later time by the association of undeformed aplites and pegmatites, and calls the process "piezo-crystallization" (see the criteria for distinguishing igneous from metamorphic foliation, page 351).

Leith also calls attention to a diagenetic flow of plastic sediments, which must be distinguished from metamorphic rock flowage.¹

How Rocks Flow.—Solids can be made to flow, acting in that one respect like liquids, by pressure that overcomes the rigidity without diminishing it. There should be no implication of fluidity in this idea. The rocks do not become liquid or gaseous during this flowage or even malleable like some metals.

When stress applied to a rock is said to deform it "within the elastic limit," this means that when the pressure is released the rock returns to its original form and size. Greater stresses deform the mass permanently by clastic, plastic or blastic processes (pages 390 and 391). The blastic process, recrystallization, and the clastic, granulation, are discussed in more detail on pages 404 to 413. Recrystallization often largely dominates, and the emphasis is placed on factors favoring rapid solution. Most metamorphism probably involves all these processes.

¹ LEITH, C. K., "Structural Geology," rev. ed., p. 228, 1923.

Much study has been given the flow of glacial ice as probably illustrating the processes of flowage in other rocks.¹ The fundamental cause of flow must operate below the melting temperature, and the mass is essentially crystalline all the time. The crystals are not oriented, but in some places there is gliding along a platy structure in a crystal (plastic). The motion is probably mostly a result of grains sliding over each other till they interlock, and in part a result of flow by melting and recrystallization (blastic) under locally intense pressure and tension; just such a process as occurs when quartz sand becomes quartzite. Near the end of a glacier there is considerable thrusting along fractures (elastic).

Where Rocks Flow.—The visible deformation of rocks near the surface of the earth is mostly by fracturing and only in very weak rocks, such as clays, by flow. Thin slabs, like window caps in old buildings may bend a

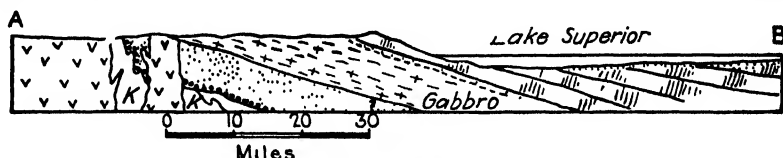


FIG. 247.—Geologic section from Lake Superior northwest. K, Keewatin; stippled areas, sediments of the early pre-Cambrian. The sediments to the left of the layer marked "gabbro" must have been more than 10 miles below the surface when the gabbro was injected, yet they are only slightly metamorphosed unless close to an igneous rock.

little after a long time, but the results of such slight deformation are not classed as metamorphic.

Erosion exposes rocks that have been deformed at greater depths, and the structures permit a fairly accurate estimate of their position at the time of flowage. At moderate and great depths in the crust, competent rocks yield elastically up to the elastic limit and then fracture; and weaker ones recrystallize and flow. The proportion of flow to fracture is irregular but in general increases with depth. The evidence of shortening found in mountain folds and thrusts shows that the deeper rocks must also have been deformed, but the absence of much fracturing indicates the dominance of recrystallization. The conditions known to favor rock flowage as distinct from fracture (page 404) also suggest that flowage occurs chiefly at great depths in the earth.

As a matter of observation nearly all the rocks that we know at the surface, some of which in the past have been more than 10 miles deep (Fig. 247), show signs of both fracture and flow. In many rocks, however, rock flowage has largely dominated the deformation. Fractures may have occurred before and after the period of flowage, but at the time of most intense metamorphism there was little fracturing. The intricate deformation of rocks at some granite contacts indicates conditions of practically pure rock flowage.

¹ HAWKES, L., The structure and flow of ice, *Geol. Mag.*, vol. 67, pp. 111-123, 1930.

Van Hise estimated that rocks would flow rather than fracture at a depth of about 6 miles, but he realized that a confining pressure might so strengthen the rocks as to increase this depth. Leith would extrapolate the observed zone of combined fracture and flow to great depths and questions whether there really is a zone of flowage. His criteria for the zone of flow are abandoned.¹ This is an extreme view.

Earthquake data are suggestive in this connection. It is claimed that 90 per cent of the earthquakes, felt at the surface, originate at depths of less than 8 km. (5 miles), and most of the rest at less than 30 km.² These disturbances at shallow depths (episeisms) are possibly to be contrasted with some of deeper origin (bathyseisms), but the method of estimating the depth and nature of the disturbance becomes less certain at depths beyond 30 km., and there may be shocks due to some disturbance that is not a fracture. It is safe to estimate that more than 99 per cent of the fractures occur within 30 km. of the surface, and that those constituting the fraction of one per cent at greater depths rapidly decrease with depth until they are negligible below 200 km.

As noted on page 170, the extrapolation of the batholiths of any period, if followed rigidly to its logical conclusion, might indicate a magma zone all around the earth. Between the idea of a magma zone and that of a zone of fracture at all depths, the idea of a zone of flow, or an asthenosphere, is to be preferred.

// It is, of course, frankly admitted that the "zone of flow" is an inference and can not be observed or tested. It is none the less a useful idea in explaining gravitative adjustments at considerable depths. The depths are so great that the rocks involved in the undertow are not exposed by later erosion—so great that conditions may be uniform over wide areas or even throughout a real shell or sphere of the earth.

In considerable contrast with these probable widespread conditions of rock flowage at depth, the same conditions as to heat, stress, solution, etc., are found locally near the surface of the earth in the roofs and wall rocks of batholithic magmas. If this view is correct, the asthenosphere or "zone of flow" is a true zone only at depths of many kilometers below sea level, but the conditions of that zone rise locally into the earth's crust and facilitate a similar flowage in a region rather than a zone.

Rock flowage no doubt occurs also in places outside of these two main regions—the deep zone and the roof and walls of magmas, for we find folded rocks at the surface far from igneous masses. Pressure in mountain folds may be great enough to deform rocks even when cold and dry, and some rocks like shales are so weak that no very great thrust is needed to deform them. The metamorphism accompanying such deformation, however, is usually slight (see Fig. 247). (The main examples of metamorphism by rock flowage are found near batholiths or in places where it may reasonably be inferred that a batholith lies close below.

¹ "Structural Geology," 1st ed., p. 111, 1913.

² OLDEHAM, R. D., Depth and character of earthquake origins, Jour. Geology, vol. 34, p. 394, 1926. F. E. WRIGHT, Scientific Monthly, vol. 24, p. 45, 1927.

✓ **Forces and Agents Favoring Rock Flowage.**—Clearly the dominant factor favoring rock flow is a stress. The secondary factors which complicate the process are heat and solutions that would favor recrystallization (see pages 406 to 409). The weakening of rocks by heat is noted on page 390. ✓

There is some question whether deep burial under later rocks may cause rock flow by simple load. We do not know with certainty of any case in which schistosity has developed without deformation. Uniform pressure is not only *not* a cause of flowage, but it increases the strength of rocks so that they resist the stresses that cause flow.

✓ **Kinds of Rock That Favor Rock Flowage.**—Many cases have been observed in which two rocks have been deformed and one flowed but the other fractured. A considerable series might be compiled showing the relative ease of flowage. Marble is probably about as susceptible as any common rock, but the clay rocks, slate and schist, are abundant and flow easily. Cherts, quartzites and granitoid igneous rocks are resistant or "competent" in comparison with shales, which are generally "incompetent" and flow freely, filling up any irregularities among fractured blocks of competent rock. ✓

Time Involved and Rate of Flow.—The rate of rock flowage is not estimated with accuracy in most places. The rate of motion of glacial ice is about one foot or more per day and may be taken as a fair indication of the maximum rate of rock flowage underground. This rate is probably rarely attained except in contact action, where heat, pressure and mineralizers are all particularly favorable. Isostatic adjustments probably seldom involve the flow of rocks at as high a rate as the rate of accumulation of sediments. Orogenic movements that involve an uplift of a few feet in a century are probably in the nature of great "revolutions," between which movements would be slower. Recent data from California¹ indicate a shift of perhaps 20 feet in 30 years.

The time involved in flow by granulation is less than that needed to recrystallize the same rock, for if it could yield by recrystallization the stress would be relieved before crystals broke up.

Results.—In a large way rock flow relieves the stresses that cause it. Petrographically rock flowage makes schists, gneisses and other metamorphic rocks. The process involves certain physical and chemical changes, which are noted in detail below as results of granulation and recrystallization. ✓

RECRYSTALLIZATION

Recrystallization is the change in texture or minerals produced as material is lost from the boundaries of some crystals and added to others.

¹ WILLIS, BAILEY, The Santa Barbara earthquake, Sci. News, July 10, p. x, 1926.

In rocks it is commonly believed to involve the solution or fusion of some crystals with the practically simultaneous growth of other crystals, making it possible for the rock to retain its general solidity. There may, however, be some recrystallization without the intervention of a liquid. Recrystallization occurs a little at a time and pervasively throughout the mass. It is essentially constructive, and minerals are commonly enlarged or new ones are formed by it. The new crystals may be of the same composition as the original or may be products of reaction. No change in the total composition of the rock is necessitated, but heat and water are so universal that some changes in hydration are common and more or less addition or loss of other materials is not rare.

Recrystallization is one of the most prominent of metamorphic processes, but the term is not synonymous with metamorphism. The recrystallization that occurs at a late stage of magmatic action is called deuteric, and the recrystallization of a freshly deposited sediment is called diagenetic, rather than metamorphic. On the other hand, metamorphism by granulation may be very intense without any great amount of recrystallization. Neither is recrystallization synonymous with rock flowage. Extensive recrystallization may occur in a rock with no appreciable flowage or deformation of the mass, as in argillites and contact metamorphic rocks; and some rock flow involves granulation with hardly any recrystallization. (It is clear, however, that the most prominent cases of recrystallization of rocks are those involved in rock flowage and producing metamorphic rocks.)

The extent of recrystallization in a rock may vary from a trace to 100 per cent. Complete recrystallization is illustrated by impure clays that are wholly changed to biotite, quartz and feldspar; also by basalt with no hornblende, which has recrystallized to a schist consisting wholly of hornblende. —

Where and When Rocks Recrystallize.—The regions in which recrystallization occurs range from the surface, where diagenesis affects newly deposited sediments, down to the greatest depths. The conditions favoring the process suggest that it would be especially active at great depths. In the assumed asthenosphere, for example, recrystallization would largely dominate over granulation and fracturing, and would rapidly adjust any disturbance of equilibrium.¹

These conditions characteristic of the asthenosphere are not known to have prevailed at any geologic epoch, near enough to the surface of the earth, so that a zone of rocks all around the earth has been uniformly recrystallized and later been exposed by erosion. On the contrary, the metamorphism of the rocks to which we have access has been local and periodic. It seems certain that the conditions of the asthenosphere have

¹ Van Hise, in "Treatise on metamorphism" p. 772, shows that recrystallization takes less energy than granulation.

been brought up, possibly within a few thousand feet of the surface, by magmas rising into the lithosphere.

Perhaps the strongest argument for the shallow depth of conditions permitting recrystallization is based on the estimates of the depth of average erosion since the ocean was formed, and of salt that was supplied to the ocean from the eroded rock.¹ An average erosion of less than two miles of igneous rock from the continents would supply the salt of the sea. As some Archean rocks are probably eroded much more deeply than two miles, others are probably eroded much less than one mile; they may never have been more than a mile below the surface of the earth. In spite of this moderate depth of burial, however, practically *all* the Archean flows are thoroughly recrystallized. This might suggest that the outer part of the earth was generally hotter in Archean time than later, and it is not altogether improbable that before the formation of the oldest rocks now known there may have been a molten surface zone of the earth or a molten zone below the surface. Radioactivity indicates that periodically through geologic time high temperatures may have developed over wide areas close to the surface, favoring recrystallization in zones so near the surface that the altered rocks might later be exposed by erosion. No rock structures or exposures, however, have given convincing evidence of the shallowness of such widespread high temperature zones. Archean time was long and the recrystallized rocks have been invaded nearly everywhere at one time or another by batholiths. It is thus possible to explain the recrystallization without any appeal to conditions differing from those of later geologic epochs.

Certain distinctions in the type of recrystallization may be based on the conditions associated with the rise of magmas in different regions. In mountain folds batholiths bring not only heat and emanations but also stresses. Stresses are much less prominent around plateau batholiths, and contact recrystallization may there be much more static than dynamic.

All this discussion has favored locations near igneous intrusives for recrystallization, but attention was called at the start to some examples in other locations. Any rock in a region of intense stresses or even of heavy load is likely to be recrystallized; and rocks like limestone may recrystallize in almost any zone or location in the earth.

Factors of Recrystallization.—The factors favoring recrystallization are chiefly water and heat, this association suggesting at once that an especially favorable agent would be igneous emanations. Some chemical forces take a prominent part, and stresses influence the result materially.

When all these factors are present, recrystallization is rapid in almost any kind of rock. When only one is prominent, the effects vary with

¹ CLARKE, F. W., *Data of Geochemistry*, 5th ed., U. S. Geol. Survey Bull. 770, p. 31, 1924. JOSEPH BARRELL, *Igneous invasion and regional metamorphism*, *Am. Jour. Sci.*, 5th ser., vol. 1, p. 185, 1921.

the activity of that factor. In diagenesis water may cause a rather extensive recrystallization at ordinary temperatures and low pressures, if the minerals are fairly soluble and exposed for a long time. Again, if a rock is heated, even in the absence of much water or stress, it may wholly recrystallize; as in the formation of hornfels from inclusions in magma. These are extreme cases, and most metamorphism probably involves both water and more heat than is normal at the surface of the ground. In fact even the production of a hornfels can not be said to occur without water, for most inclusions in magmas have hydrous minerals, and practically all magmas contain a little water. It should be noted, also, that only a little water is needed to produce a large effect, for the process of recrystallization does not use it up; a little water may be used repeatedly to recrystallize successive small fractions of the mass until in time the whole rock is changed.

The factors favoring recrystallization have received intensive study by metallurgists on account of the common recrystallization of metals and alloys, and the resulting changes in the properties of the metals. Certain of their conclusions¹ may be of value here, because their experiments involve neither water solution nor fusion.

Grain size in cast metals which undergo no phase change in the solid state cannot be changed appreciably by heating below the melting point.

Grain growth in the solid state may occur in compressed metal powders and in alloys in which a new phase is formed in the solid state.

The recrystallization temperature is lower (a) the greater the amount of previous deformation, (b) the smaller the grains prior to deformation, (c) the lower the temperature at which deformation is effected, and (d) the longer the time of heating.

In any given metal the grain size, after complete recrystallization, is normally smaller the lower the recrystallization temperature.

Although the dependence of recrystallization on water solutions is not complete, there are several examples of metamorphism once attributed to dry reaction that are now known to be dependent not only on water but on water with dissolved mineral matter. The formation of calcium silicates in sandy limestone, for example, results from introduced solutions of silica at much lower temperatures than that required for the recrystallization of materials already there (page 422).

— Pressure effects are subordinate to heat effects in general, and there are several noteworthy illustrations. Direct uniform pressures usually raise the melting point of a substance, but 1,000 atmospheres are not so effective as 40° change of temperature. Similarly, in the few cases tested, the solubility of substances is affected less than 5 per cent by a pressure of 1,000 atmospheres,² though high temperatures may make a silicate miscible with water in all proportions. Direct pressure may also cause

¹ JEFFRIES and ARCHER, *op. cit.*, pp. 117–118, 1924.

² JOHNSTON and NIGGLI, Principles underlying metamorphic processes, *Jour. Geology*, vol. 21, pp. 504–505, 1913.

some transformations involving the density of minerals. Recall the curves of the several forms of ice (page 153).

Indirectly pressure influences solution very greatly by holding gases like CO_2 in solution and by keeping water in a liquid form at high temperatures.

- Field observations indicate that the pressure of a load of overlying rock a few thousand feet thick may cause some recrystallization with little deformation (see page 398). Great areas are known where rocks show no sign of deformation, but have a horizontal schistosity.¹ Such structures are rare, however, and probably indicate that the lateral support was weak or even replaced by a tension of some sort. If such a schistosity is not present in the altered rock, there is always doubt whether the recrystallization was due to pressure; it may have been heat. Nevertheless some shales that do not seem to have been much heated have chlorite and mica plates developed in haphazard portions by recrystallization. The relative importance of temperature, pressure and water is not determinable in such rocks.

- The effects of stress or non-uniform pressure in recrystallization are much more important than those of uniform pressure but still subordinate to those of temperature and water. Stress favors solution and by a selective solution of the strained portion of crystals causes the orientation of elongated mineral grains. The strain is greatest where the grain is pressed by adjacent grains and least at right angles to that pressure. The grains dissolve at the points where they are strained and grow on the sides where not so much strained (Fig. 246). Some flat grains in schists may be largely made up of original mineral material with additions on the sides.

A further but more obscure method of development of parallelism of plates is indicated by the results—a schistosity in which the plates of mica are not only elongated in a plane, but the main platy minerals show more or less crystallographic parallelism (page 400).

Chemical reaction is prominent during recrystallization because the materials pass through solution. Quite aside from the additions, subtractions and exchanges of material from a distance, the constituents of a rock often recombine to new chemical molecules without notable change in rock composition. The basis of reaction is chemical, but the minerals formed vary not only with the concentration of material in the system but with the temperature and pressure (pages 391 to 400). A few of the simpler reactions such as the reversible reactions of hydration and dehydration are fairly well understood. It is also clear that in a magnesian limestone with silica, the magnesia combines with silica before the lime does (see further the diagrams of suggested equilibria). Rather

¹ DALY, R. A., *Metamorphism and its phases*, Geol. Soc. America Bull., vol. 23, p. 400, 1917. C. K. LEITCH, *Rock cleavage*, U. S. Geol. Survey Bull. 239, pp. 68-71, 1905. ARTHUR KEITH, U. S. Geol. Survey, Folio 124, p. 7.

more of the reactions, however, are complicated by the large number of components involved, the complex nature of the silicates formed and the lack of many data about their composition. There has even been a suggestion that the composition of the whole rock might be somewhat modified by the development of minerals with a limited range of composition. This is a "convergence to type" in metamorphism (see pages 420 and 421).

The kind of rock to be recrystallized has a great influence on the process. The chief examples of easy and rapid recrystallization are the



FIG. 248.—Sketch of metacrysts of silicates in a thin section of calcareous schist, Rhodesia. The included biotite and sericite are oriented so as to indicate that the metacrysts around them formed by recrystallization, and the material of the inclusions was not needed in the formation of metacryst, *helizitic structure*, $\times 15$.

calcite rocks. Marble is almost invariably so recrystallized as to obscure any effects of deformation, which in some cases it is known to have undergone. This is believed to be a result of the great solubility of the calcite in comparison with that of other minerals. Some less common carbonates have a similar tendency to recrystallize. Other features of a rock that seem to favor recrystallization are a porosity sufficient to allow some access of solutions and a complexity of composition such that chemical reactions are likely to take place. These factors may be illustrated by the nature of a common clay, with many minute pores and a complex of constituents, which by reaction readily form some member of the mica group. The beginning of this recrystallization to mica does not wait for much heat or pressure in the clay.

Selective Recrystallization and a Sequence of Effects.—The conditions thus outlined for recrystallization may develop in various combinations and sequences. The results vary with the factors involved, and after a long sequence of changes the late effects obscure the early ones. It is therefore important to note that recrystallization may be selective, some minerals in a complex rock being changed while others are little affected (Fig. 248). In this partial recrystallization we find clues for determining the original of some altered rocks and can apply them to other rocks, even to those that are wholly recrystallized.

In general mica recrystallizes early and hornblende almost if not quite as early. Carbonates may precede all the silicates. Quartz comes a little later than mica, and feldspar still later. Metacrysts are later than their matrix, but some metacrysts may have formed early only to be destroyed at the period of most intense deformation of the matrix. The sequences reported in several districts are almost wholly based on sedimentary original rocks, so that the earlier effects are those at shallow depths and the later effects those of the deep zone (see pages 357 and 432 and the readings on zones of metamorphic rock). It is clear, however, that rocks originating under deep igneous conditions would show a different sequence, the minerals of the deep zone being followed by those of the shallow zone, if deformation continued while the rock was gradually brought nearer to the surface.

Time Required and Rate of Recrystallization.—From the rapid diagenesis of some limestones, which is practically complete before the next bed is deposited, to the slow change of early pre-Cambrian slates, which are little altered, there is a wide range in the rate of recrystallization. In many silicate rocks the great accelerators of the process are heat and water, especially the heat of igneous intrusives near by; but even in this case the process must require a certain amount of time, for many intrusives show a surprising lack of metamorphic action.

Results of Recrystallization.—The general results of recrystallization may be summarized on the basis of kinetic equilibrium. (1) Small crystals disappear and large ones grow larger; thus slate becomes schist, and deep-seated rocks approach granitoid textures. (2) Strained crystals dissolve and unstrained crystals grow; thus a marble intensely folded soon loses the signs of strained crystals. (3) Those portions of grains that are free from strains grow at the expense of strained portions; thus a sandstone under stress forms a quartzite with oriented quartzes and with little sign of strain. (4) Crystals of the right orientation grow, and others disappear or fail to grow; thus a biotite schist has grains orientated optically and in mineral cleavage as well as dimensionally. (5) Unstable crystals dissolve, and stable minerals form; thus aragonite becomes calcite. (6) Pore space is reduced especially in sediments. (7) Much recrystallization under pressure forms minerals of greater density, at least if the original is sedimentary. (8) Much recrystallization is accompanied by some change of composition.

The more specific results of recrystallization are the metamorphic rocks and minerals (pages 434 to 436), and a few other rocks like recrystallized limestones. There is a notable development of platy minerals—micas, amphiboles, and chlorites (for the relative dimensions see page 383). These give the rocks their cleavage; but some rock cleavage may occur in other rocks when sheared or when the parallelism is a primary feature instead of a recrystallization. It is believed that in some

rocks recrystallization may produce a segregation of material into bands.

The criteria by which recrystallization may be recognized are here tabulated and form a good summary of results.

CRITERIA SUGGESTING RECRYSTALLIZATION

Mineral criteria.

1. Chlorite, grunerite, tremolite, actinolite, garnet, staurolite, kyanite, chloritoid, tourmaline, magnetite, cordierite, vesuvianite, rutile, epidote, sillimanite, andalusite, zoisite, ottrelite, wollastonite, mica, hornblende, sericite.
2. Quartz showing elongation, secondary growth, phantom veinlets, dovetailing of crystals and ribboned structure.
3. Mica that feathers out against adjacent crystals. Bounding of several quartzes by a biotite leaf or flake.
4. Metacrysts of minerals (not igneous phenocrysts), garnet, staurolite, kyanite, ottrelite, etc.
 - a. If inclusions in a large crystal are mostly quartz and groundmass is quartz feldspar and ferromagnesian minerals, they are metacrysts.
 - b. Presence of crystals with inclusions that are related in size, shape, position and composition to minerals of groundmass.
5. Bleached and discolored zones around mineral core; aureoles.
6. Minerals that could not be primary; *e.g.*, garnet and tremolite in a limestone; chlorite or sericite in granite; pyrite in a sandstone, etc. (*if not introduced*).
7. Soluble minerals such as carbonates.
8. Presence of large dense crystals.

Textures and structures.

1. Pseudomorphs and pseudomorphous textures, except weathering and minerals formed wholly by replacement.
2. Sutured, mosaic, sugary, poikiloblastic, porphyroblastic, or granoblastic textures. Absence of wavy extinction in deformed rock. Secondary growths.
3. Schists. Foliated and banded structures due to parallelism of platy minerals such as mica, chlorite, etc. Feathering out of micas. *Lil-par-lil* injection.
4. Evenness of grain, coarseness of grain or increased size of grain; low porosity.
5. Perlitic structure in crystalline rock.
6. Different texture or order of crystallization from that expected of igneous minerals (pages 44 and 357).
7. Occurrence of minerals with their greater diameters at right angles to the original structure, such as bedding, which do not show sufficient deformation to justify the assumption that they reached their present position by rotation.
8. Shape of mineral particles in deformed rock as compared with the shape of the same minerals in primary rocks. Ratio of elongation. hornblende more than 5:1; mica and chlorite more than 6:1.
9. Segregation of like minerals into bands.
10. Preservation of schistosity by included minerals.
11. A recrystallized mineral may have more perfect crystal outline than its inclusions, opposite in an igneous rock.

Other criteria.

- Field evidences of depth, high temperature, long time, water.
- All criteria for contact action.
- Most signs of deformation suggest recrystallization.
- The presence of igneous rock near by.
- Favorable nature of the rock.
- In a series of specimens, changes in grain size, changes in color, in number of inclusions, increase in per cent of minerals of higher density.
- Strength in a granulated rock.
- Lack of fragmental textures, mortar structures, bent and broken grains, strain effects.
- Lack of skeleton crystals; and zonal growth; and glass.
- Lack of inclusions in quartz.
- Absence of tridymite and marcasite.

GRANULATION

Granulation is the crushing of a rock under such conditions that no visible openings result. The fine fractures extend through the minerals, distinguishing granulation from the coarser structures of faulting and jointing. The grains may be reduced to thousands of fragments, but under a confining pressure the minute fragments do not lose their coherence, and the rock remains almost as strong as before. In contrast to recrystallization, granulation is wholly a matter of rock flowage or deformation.

Laboratory experiments in rock flowage are accomplished largely by granulation. Adams'¹ work indicates that below some 30 miles the openings in rocks are not likely to persist. Deformation at greater depths is by granulation or recrystallization rather than by open fractures. For some rocks the depth of granulation may be much less than for others.

The factors favoring granulation are largely contrasted with those favoring recrystallization. First and above all others, intense and rapidly applied stress is the dominant factor. The time involved in the deformation must be short, for rocks seem to be deformed without granulation if the stress is moderate but prolonged. A confining pressure usually related to depth is also essential. Finally, the minerals of the rock deformed have a good deal to do with the matter; brittle, hard, insoluble minerals like quartz and feldspar are granulated more than soluble, soft minerals like gypsum. A limestone may be granulated but recrystallizes so readily that the signs of granulation are soon almost lost.

In contrast to the conditions of batholithic roofs where rocks recrystallize, rocks are granulated in zones of deformation farther from igneous intrusives. Both processes may be characteristic of mountain folds, but the granulation is more prominent in the colder, shallower, and drier portions of the folded structure.

Granulation always reduces the grain sizes, forming successively mortar, augen, and mylonite structures. Usually if there are grains of different colors, the rocks become banded by a flattening of each grain into a platy or lenticular aggregate. There is clear evidence in some rocks that broken slices may be somewhat rotated also into the plane of the new structure (Fig. 218). The banding and orientation give the rock a certain degree of cleavage, but not so perfect a cleavage as often develops by recrystallization. Certain minerals may be said to be characteristic of granulated rocks, but probably if they are formed during metamorphism they resulted from recrystallization. Mylonites

¹ ADAMS, F. D., Depth of the zone of flow, Jour. Geology, vol. 20, pp. 97-118, 1912. — and J. A. BANCROFT, Internal friction in rocks, Jour. Geology, vol. 25, pp. 597-637, 1917.

are characteristically feldspar and quartz. It is an unsettled question whether crushing in rocks already warm may develop heat enough to produce actual magma or glassy rocks. Some rocks that appear to have formed by crushing are isotropic under the microscope—pseudotachylite.¹

OTHER METAMORPHIC PROCESSES

Plastic Deformation.—Plastic deformation is in a sense intermediate between recrystallization and granulation. Moderate stresses deform cleavable minerals by gliding and development of twinning; they may bend the flexible minerals like mica and develop strain shadows or wavy extinction in most crystals.



FIG. 249.—A migmatite or mixed gneiss, in which it is clear that some bands are of igneous origin and are connected with a dike. See also Fig. 46.

Introduction.—Introduction of material may be prominent in some cases but very slight in others. The examples of prominent introduction include the injection gneisses, in which igneous material follows the structure of a schist; and a number of rocks in which a pure limestone has become garnet rock by contact action; and some rocks hydrothermally altered in the walls of veins. In a metamorphosed rock of complex composition additions of material are not always easy to distinguish from simple recrystallization. (See further readings on Introduction.)

Lit-par-lit injection is so clearly a matter of magmatic intrusion that there has been little question in recent years about the reality of the mixed gneisses (Fig. 249) part schist and part granite or some granitoid igneous rock. Such rocks are abundant, and because of their banding they are classed as metamorphic, even if they contain a high proportion of igneous material. The exact conditions of invasion are not so certain. The thin leaves of schist between injected dikelets are in many cases undeformed, showing that no very viscous material was violently thrust in. Probably the injection may be compared to the wetting of a pad of

¹ SHAND, S. J., Pseudotachylites of Parys, Geol. Soc. London, Quart. Jour., vol. 72, pp. 198-221, 1916.

paper held so as to dip into a dish of water. Capillarity may draw water up between the leaves and even soak into the paper leaves themselves. The magma, however, is of complex composition as compared with water, and the *lit-par-lit* injections probably constitute an "advance guard" of the magma, more dilute and aqueous or at least somewhat different from the main magma. They may possibly be gaseous or watery separates from the magma. Some late magmatic residues also inject schists (Fig. 262).

Sederholm in particular uses the term "granitization" for a series of effects due to granites, some of them, however, not making the wall rock approach very closely to the nature of granite. The wall rocks with igneous additions are "migmatites," and the process of digestion of wall rock into magma is "syntexis." The extreme of metamorphism without addition, Sederholm believes to be fusion to a magma (page 230). "Ptygmatic" complex folds (Fig. 224) are common near injection gneisses and schists and suggest to many that the wall rocks of magmas may become "mushy" or viscous or even half fused.

Pervasive emanation and contact action resemble *lit-par-lit* action, but as the fluids are not introduced into a schist they do not follow layers. The high temperature may so reduce the viscosity of the emanations that they pervade a rock in submicroscopic channels. The temperature is probably above 374° , the critical temperature of water. Mineralizers are probably as important as heat in contact action, carrying not only heat but also active chemical solvents and reagents. Even where *lit-par-lit* injections occur, pervasive emanations may extend far beyond the zone of recognized igneous material.

The materials introduced in contact action and their sequence have been noted under igneous emanations (pages 216 to 220). Silica and iron come early, oxides later, and sulphides still later. The proof of their introduction has been largely the work of economic geologists. They have in some rocks been able to find structures indicating a certain origin and a mineral composition that would be destroyed by such an origin. The structure may be large, as in tracing a sedimentary formation; or small, as in fossils that are replaced by pyrite or garnet. (See the list of further readings.)

Farther from the magma emanations play a large part in the formation of ore deposits and cause hydrothermal metamorphism in the walls of veins. This involves both addition and subtraction (see metasomatism below).

It is much less certain that any considerable quantity of material is added during regional metamorphism. Most micas, however, contain fluorine, and its abundance in a regionally metamorphosed schist suggests the introduction of the fluorine, which is not found so commonly in clays from which the mica schists are formed. Evidence of addition is even stronger where tourmaline, containing both fluorine and boron, is abundant over a considerable region or where ores are introduced.

Subtraction.—Subtraction or leaching during metamorphism is less often noted. It is well understood, however, that hydrous rocks like clay lose water and that impure carbonate rocks lose CO_2 . Locally the silica

may be leached from a sericitic quartzite so completely as to leave a sericite schist with only traces of quartz.¹

Metasomatism or Replacement.—Replacement is a term used by petrographers in a rather technical sense, and as it is also a term used popularly in a broader sense, the technical term *metasomatism*² is preferable. Metasomatism or replacement is a practically simultaneous solu-

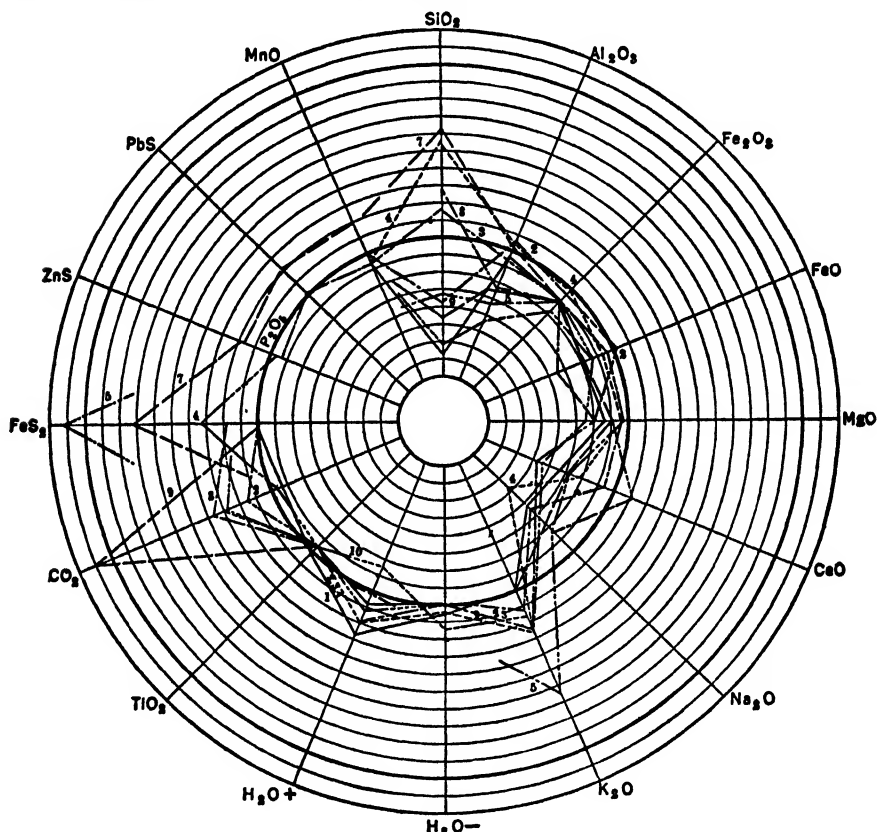


FIG. 250.—Sericitization and propylitization. The heavy circle represents the composition of the unaltered rocks. Losses in hydrothermal alteration are measured toward the center and gains are measured outward. Each broken line threading the spokes of the wheel represents one instance of rock alteration. (After Reber, *Econ. Geology*, vol. 11, p. 566, 1916.)

tion and deposition through small openings, usually submicroscopic, and mainly by hypogene water solutions, by which a new mineral of partly or wholly differing composition may grow in the body of an old mineral or mineral aggregate. The limitation to hypogene water solutions excludes

¹ LEITH, C. K., and W. J. MEAD, "Metamorphic Geology," p. 126, Henry Holt & Company, New York, 1915.

² LINDGREN, WALDEMAR, *Metasomatism*, Geol. Soc. America Bull., vol. 36, pp. 247-248, 1925.

from metasomatism some of the common pseudomorphs resulting from weathering and some cases where magma replaces a rock by assimilation.

As the term metasomatism is used by Goldschmidt it means a change in rock composition rather than in minerals. He notes several main types,¹ but the number is legion.

The common examples of metasomatism in metamorphism are those of contact action and those of hydrothermal attack on the walls of veins. The prominent effects noted in contact action are the replacement of limestone by andradite and other heavy silicates, the replacement by iron ores, the formation of metacrysts in schists (Fig. 248), and a general silification of both limestones and shales. Hydrothermal action typically propylitizes and sericitizes the igneous rocks. This action has been



FIG. 251.—Zoned garnet and calcite in a contact metamorphic rock, Golconda, Mexico. Crossed nicols. The garnet during replacement of a limestone developed euhedral forms, and the calcite (light with twinning bands) fills in around it and partly replaces alternate zones in the garnet. $\times 30$.

studied chemically as well as mineralogically² (Fig. 250). Potash, sulphur, and water are added; soda, lime, etc., are removed. It is very remarkable that with such important changes in chemical constituents the volume may remain almost constant.³ Some hydrothermal carbonation commonly follows contact action in a limestone (Fig. 251).

¹ GOLDSCHMIDT, V. M., On metasomatic processes in silicate rocks, *Econ. Geology*, vol. 17, p. 106, 1922.

² STEIDTMAN, Alteration of rocks by weathering and hot solutions, *Econ. Geology*, vol. 3, pp. 381-409, 1908. REBER, Mineralization at Clifton-Morenci, *Econ. Geology*, vol. 11, pp. 528-573, 1916.

³ LINDGREN, WALDEMAR, Contact metamorphism at Bingham, Utah, *Geol. Soc. America Bull.*, vol. 35, pp. 507-534, 1924. JOSEPH BARRELL, Physical effects of contact metamorphism, *Amer. Jour. Sci.*, 4th ser., vol. 13, pp. 279-296, 1902.

It is noteworthy, also, that some minerals are replaced by colloidal gels.¹ Many of these are in the shallow zone, are results of diagenesis,



FIG. 252.—Chalcocite and other secondary minerals replacing primary pyrite, Chuquicamata. The cores of pyrite (py) are irregular, but the replacing minerals are in zones and, part of them being transparent, they suggest that the mechanism of replacement was by the solution of a thin layer and the filling of that space by new minerals. $\times 100$.



FIG. 253.—Sketch of thin section of the Rand conglomerate, South Africa. Pyrite crystals transgress upon the outlines of original grains, showing some replacement of the quartz. $\times 10$.

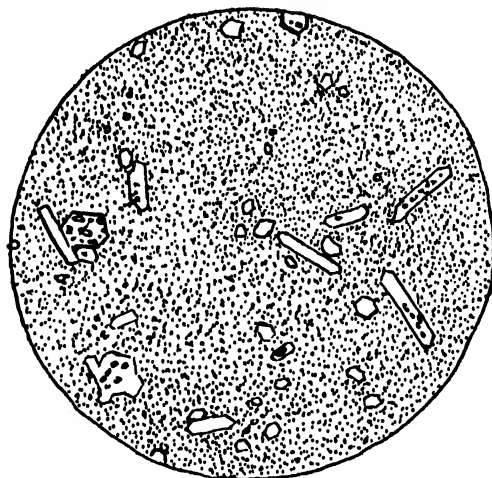


FIG. 254.—Incipient silicification of a limestone, Aspen, Colorado. Euhedral quartz crystals are clear or include small grains of limestone. $\times 30$. (After Lindgren, *Trans. Am. Inst. Min. Eng.*, vol. 30, p. 628, 1900.)

and are formed by cold water; but some deeper deposits seem to have passed through a stage of hydrothermal gel replacement. Many pseudo-amygdules in basic lavas have structures suggesting gel replacement.

¹ LINDGREN, WALDEMAR, *Geol. Soc. America Bull.*, vol. 36, p. 253 *et seq.*, 1925.

As the process of metasomatism seems to proceed volume for volume, chemical equations and the law of mass action do not wholly control. The phase rule does not apply, at least to gel replacements. Some such replacements may mean solution and filling of cavities of appreciable size, but probably not all are of this type (Fig. 252). The textures and structures of the original are so well preserved in many rocks that openings are probably submicroscopic. Material is transferred by flow, part of it capillary flow, and by diffusion.

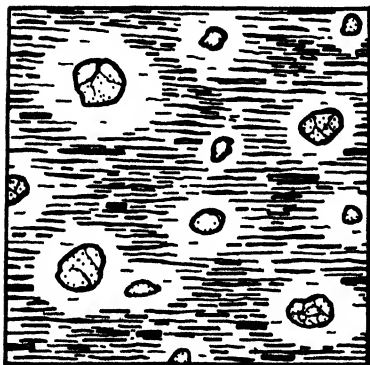


FIG. 255.—Garnet metacrysts growing in somewhat imperfect forms by recrystallization of a hornblende schist, from Georgia. The garnet uses the iron of the nearby hornblende, leaving a bleached zone around the garnet in the schist. About natural size.

Replacement may be partial or complete. When magnetite is replaced by pyrite some of the iron of the original may be used in the new mineral. On the other hand, when a limestone is silicified, the replacement may be so complete that no relic of the original material is found in the altered rock.

The results of metasomatism are chiefly new minerals, textures usually being remarkably preserved. Some of the altered rocks may be porous, others are compact. There is no good term to describe the relict texture

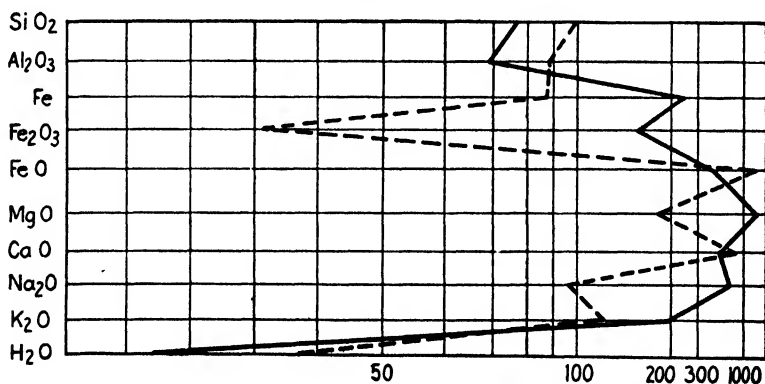


FIG. 256.—Straight-line diagram showing gains and losses of constituents of diabase and basalt by weathering, from an original taken as 100. In this diagram the several constituents are each given a horizontal line, and one pair of rocks (fresh and altered) gives a series of points connected by a zigzag line. (After Leith and Mead.)

of a rock in which minerals are replaced. It has been called a "pseudomorphic texture." Each replaced grain, however, in a granitoid rock is a "pseudomorph" of a grain that did not have a crystal form in the original but only the irregular form of crystals that mutually interfered.

The criteria of recrystallization and replacement are complex and are given in some detail (see page 411). The processes are probably so prolonged that the chemical systems normally attain an equilibrium at the high temperatures reached. Of course, the cooling that follows the metamorphism of most rocks leaves the minerals in a condition tending to further change, but this may be very slow.

The chemical and quantitative studies of metamorphism that involve changes of material are very difficult unless we know that the volume, or some certain constituent, remains constant from the original to the altered rock (see Problem 8, pages 20 and 21). The "straight-line" diagram gives a good method of study, if no constants can be assumed (see Fig. 256 and the further readings listed).

CRITERIA SUGGESTING METASOMATIC REPLACEMENT

I. Structures.

1. Preservation of rock and mineral structures in ores or rocks of different composition; as stratification, cross-bedding, fossils, faults, folds, joints, brecciation, dolomite rhombs, vesicles, etc.
2. Unsupported structures (except in igneous rocks).
 - a. Attitude of rocks the same in "islands" as in "mainland."
 - b. Rounded forms, etc.
 - c. Zones of minerals in inclusions.
3. Diffusion banding (an alternation of two minerals), but not crustified banding or vugs.
4. Faceted crystals intersecting original structures and grains, or present in foreign masses; *e.g.*, pyrite in shale, garnet in limestone, crystals intersecting bedding or schistosity.
5. Presence in an igneous rock of minerals not of igneous origin, as siderite, calcite, arsenopyrite.
6. Phenocrysts in an igneous rock containing inclusions of minerals commonly of later crystallization such as quartz in ferromagnesian minerals.
7. Minerals fill veins the sides of which do not match.
8. Ore related to fissures and certain types of beds with fading out of the ore at a distance.
9. Thorough silicification, but exclude pegmatites, chert, sand, etc.
10. Great variety in size and extreme irregularity of form of deposit.
11. Gradational contacts; and saw-tooth contacts.
12. Forms rounded, but not scalloped or with concave surfaces.
13. Fragments in vein oriented as in walls.
14. Relation of deposit to an impervious barrier.
15. No cave detritus.

II. Textures.

1. Pseudomorphs of minerals, fossils, rock textures, etc.
2. Penetration of mineral along cleavage or fracture planes of another as blades or like graphic intergrowths.
3. Excess of introduced minerals over porosity of original rock; *e.g.*, great excess of pyrite in sandstone.
4. Quartz shows wide variation in size of grain.
5. The habit of replacing quartz in fine rods; myrmekite and symplektite.
6. Dendritic structures in inclusions.
7. "Floating ice" as distinct from "exploding bomb" texture.
8. Mineral grains showing a fringe of another.
9. Clear crystals in a dusty rock.

III. Minerals common in replacement.

Too numerous for detail. Chlorite, sericite, serpentine, quartz, carbonate, pyrite, garnet, uraltite, epidote, fluorite, barite. The replacing mineral commonly *selects* certain minerals of a complex original rock, and attacks them before the others.

IV. Minerals as signs of replacement.

1. Association of metallic minerals with hydrothermal minerals, or even the occurrence of hydrothermal minerals; *q.v.*
2. Large proportion of minerals not primary.
3. The presence of contact metamorphic minerals such as garnet, wollastonite and pure albite, especially if a high per cent of the mass. Many other signs of contact action; *q.v.*
4. Certain mineral combinations, such as sericite and sulphides, garnet and sulphides, oxides and sulphides.
5. Presence of minerals that do not form by replacement (such as zircon, olivine, apatite, nephelite, basic plagioclase, chromite, leucite, etc.) in a matrix that is not primary to such minerals, or with a distribution and size characteristic of the country rocks, indicates replacement of the matrix.
6. Sedimentary or other rocks containing minerals that could not have been there when the rock formed. (a) Large euhedral quartz crystals in limestone. (b) Pyrite crystals in rhyolite. (c) Galena crystals in quartzite.
7. Chlorite, pyrite, etc. developed adjacent to and paralleling faults and ore bodies. Propylites.
8. Greisen, and luxullianite (deuteric replacements).

Convergence to Type in Metamorphism.—The prominent metamorphic rocks are so few and are derived from such a variety of original materials that there is a very definite suggestion of convergence to type.¹ Leith notes three prominent types resulting from regional metamorphism—sericite schist, biotite schist and hornblende schist.

It is clear that micas, hornblende and chlorite sometimes increase during recrystallization until they obscure the nature of the original. The schists analyzed are thought to depart from the original toward one or more of these minerals. In this sense there is a mineralogical convergence, and there may be some tendency for the composition of the rock to be modified because of the compositions of the minerals formed.

There is also a textural convergence due to recrystallization. If there are dynamic effects during recrystallization, the result is a schist. If not, the rock recrystallizing under static conditions usually forms a hornfels. This seems to be a result of many grains of different minerals, all growing at once, tending to produce rounded forms; less commonly some poikilitic textures are also formed. Exceptions to this textural convergence are chiefly the rocks consisting almost wholly of one mineral. A pure quartz, or pure iron oxide or pure pyroxene upon recrystallization is likely to become coarse grained, approaching a pegmatitic rather than a normal hornfels texture. Such monomineralic rocks being relatively rare, it may be said in summary that the textural convergence is chiefly to schists and hornfels.

The minerals involved in mineral convergence are such variable isomorphous series, that they accommodate a variety of originals and absorb a considerable amount of miscellaneous residue after the more-uniform minerals are crystallized. Hornblende especially can be shown to have used up nearly all the essential components

¹ LEITH, C. K., and W. J. MEAD, "Metamorphic Geology," pp. 200-201, 1915.

of a greenstone, turning into the single mineral hornblende what was originally plagioclase, augite, olivine and magnetite and at an intermediate stage had been chlorite, epidote, calcite, and quartz. Still other minerals, however, may be selected to cover the same range in composition. The fact that sericite, biotite and hornblende are formed in preference to the others shows that rock flowage finds these minerals most adaptable. Whether it is because of crystal habit, cleavage, composition or temperature is uncertain. The schists result from a combination of composition, crystal forces and environment.

This general discussion of convergence is broader than simply a discussion of schists, for contact rocks as well as regionally altered rocks show convergence. The rocks at granite contacts are usually schists because granite magmas are usually accompanied by stress. Gabbro contact rocks are usually hornfels because recrystallization occurs under more uniform pressure. Sometimes, however, granite produces hornfels, when intruded without stress. The mineralogic modifications in hornfels naturally vary with the composition of both the magma and the country rocks, but the texture, depending upon stress, differs in different masses of granite, even if the granites are of the same type.

KINDS OF METAMORPHISM

Contact Metamorphism.—Contact metamorphism is the change that occurs in the walls of an igneous intrusion or in fragments of such walls as a result of the heat and materials emanating from the magmas.¹ Internal effects on the magma, sometimes called endomorphic contact metamorphism as distinct from ordinary exomorphic effects, have been described above as assimilation or the contamination of magma (pages 225 to 231).

Contact metamorphism in many rocks involves little but recrystallization, but in others there are additions from the magma and metasomatic replacements. In clays and silicic rocks there is a tendency to convergence, producing biotite schist if there is stress, and hornfels where stress is absent.

The chief factors involved in contact metamorphism are heat and magmatic emanations. Depth and confining pressures are perhaps necessarily moderate—great enough to prevent too rapid an escape of hot solutions but not great enough to keep all the volatile elements inside the magma chamber. (See pages 210 to 216 on the physical chemistry of magmas and gases, and pages 218 to 220 as to the gaseous or liquid state of the emanation.) Depth to a certain distance means an increase in temperature and to that extent favors metamorphism. The size of the

¹ Tyrrell in his "Principles of Petrology," p. 289, distinguishes "pyrometamorphism" as hotter and drier than ordinary contact action. Some other geologists also object to "contact" metamorphism, because much of the altered material is at a little distance from a contact.

mass of magma controls the heat contribution perhaps as much as its temperature. The time required for contact action is probably short compared with that involved in regional metamorphism.

It is believed by many that dry heat causes little contact action except a baking or dehydration. Recrystallization and silicate formation are more dependent upon water solutions,¹ and it is not generally believed that heated meteoric waters have much part in contact action. The source of the solutions has been long discussed. If meteoric water was heated by a magma in contact action, the hottest magmas, basaltic or gabbroic in composition, should produce the greatest contact metamorphic deposits, but it is found, on the contrary, that most ores are related to quartz monzonite and more silicic magmas. The proof that the additions are from the magma, and not from the substances leached from the wall rocks is strongest when the inclusions of limestone inside an igneous rock are garnetized.² The waters are almost certainly magmatic in origin.

The proof of addition of silica and iron in contact action was made conclusive by classic studies at Clifton-Morenci³ and at Tamaulipas.⁴ When a limestone with very little of any constituent other than CaCO_3 becomes a mass of garnet that occupies the same space as the original, the proofs of addition of material must convince everyone.

The sequence of the effects of magmatic emanation is discussed on page 220. The contact minerals that most conclusively indicate magmatic contributions are tourmaline, axinite, scapolite and the metallic ores. Contributions of soda are shown by albitization of shales and contributions of water by the formation of hydrous silicates in a limestone.

The nature of the magma or the stage it has reached in differentiation largely determines the kind of emanation, and the nature of the wall rock determines whether the emanations will greatly alter it. Above all others the favored contacts are those between a silicic or medium granitoid igneous rock and an impure limestone.

The relative effects of different igneous rocks have not been studied in many places, but both granite and gabbro may have contact zones with a width of half a mile. The general impression that gabbro contact action is subordinate is probably due partly to the fact that gabbro is not as abundant as granite and partly to the scarcity of important ore deposits associated with gabbro contacts. It is believed, also, that gabbro magma is less hydrous but hotter than granite magma.

¹ WINCHELL, Petrographic studies of limestone alteration at Bingham, *Am. Inst. Min. Met. Eng. Trans.*, vol. 70, p. 884, 1924.

² UMPLEBY, J. B., Geology and ore deposits of the McKay region, Idaho, *U. S. Geol. Survey Prof. Paper* 97, pp. 40-79, 1917.

³ LINDGREN, WALDEMAR, The copper deposits of Clifton-Morenci, Arizona, *U. S. Geol. Survey Prof. Paper* 43, pp. 153-155, 160-164.

⁴ KEMP, J. F., Ore deposits at contacts of intrusive rocks and limestone, *Econ. Geology*, vol. 2, pp. 5-11, 1907.

CRITERIA SUGGESTING EXOMORPHIC CONTACT METAMORPHISM

Definition: The physical and chemical action of hot magmas and their products on neighboring rocks.

1. Common changes are: baking, induration, vitrification and recrystallization.
2. Contact metamorphic minerals that are very rare in any other rocks: staurolite, vesuvianite, wollastonite, birefringent garnet, cordierite, kyanite, sillimanite, andalusite, scapolite, axinite.
3. Ore minerals: *magnetite*, *chalcopyrite*, *pyrite*, *pyrrhotite*, *sphalerite*, *galena*, *molybdenite*, *specularite*, *graphite*, *corundum*, *cassiterite*, *chalcocite*, *bornite*, *ilmenite*, *chromite*, *wolfenite*, *wulfenite*, *scheelite*, *gold*, *silver*, *arsenopyrite*. The usual sequence is magnetite after silicates, and sulphides after magnetite.
4. Other gangue minerals: garnet, epidote, diopside, tremolite, zoisite, micas, hedenbergite, forsterite, fayalite, anorthite, ilvaite, topaz, actinolite, tourmaline, danburite, hornblende, spinel, rutile, apatite, quartz, feldspar, prehnite, humite, recrystallized coarse calcite, dolomite, ankerite, fluorite.
5. Rocks.
Hornfels, greisen, marble, skarn, gneiss, garnet rocks.
6. Textures and combinations.
Intimately interlocked crystals of heavy silicates, oxides and sulphides, or any two of them, practically simultaneously formed.
Greasy appearance of massive garnet and vesuvianite on fresh fracture.
Feathery texture of ore produced by actinolite and tremolite.
Greenish flaky appearance of ore due to mica and chlorite.
Poikilitic and knotty; spotted schists.
Sutured and interlocked grains. Clouds of opaque dust in feldspars.
Helizitic texture or oriented inclusions in metacrysts.
Replacement along cleavage planes and intergranular spaces.
Hornfels or granoblastic.
Lit-par-lit.
Diffusion banding, but not crustified banding.
7. Any signs of igneous emanations, recrystallization or intrusion, are suggestive.
8. Field criteria.
Within a mile, usually within one-half mile of an igneous mass of intermediate composition.
Bunchy, irregular, tabular, pipe like (oftentimes selective along fractures and in calcareous beds) in shape.
Gradually fading metamorphism toward sediments, especially toward shales. The "kindly look" of the rock, altered, etc.
Retention of bedding, and primary structure of sediments.

(See list of further readings.)

The influence of wall rock on contact action depends largely on whether or not it reacts with the igneous emanation. As the emanations are siliceous, siliceous rocks are less notably affected than the limestones, which have only a weak acid and that easily volatilized (Fig. 244). The porosity of a wall rock may also considerably affect the penetration of solutions, but at high temperatures the emanations pervade even rocks with no visible openings. Composition dominates. This is a case of chemical affinity in metamorphism.

Lime in a rock seems to act almost like a filter, catching the mineral content of a magmatic emanation, silica foremost but also iron and other earths. The percentage of lime in the rock, however, need not be high in order to have it accomplish this result. Some calcareous shales and even carbonated igneous rocks may form typical garnet zones by contact action. The dominance of calcareous rocks is emphasized by the occurrence of many valuable ore deposits in contact-metamorphosed limestones. Some contact ores are found in quartzite, but not many, and

shales and igneous rocks are less favored. The action here is more a recrystallization less an addition, and the recrystallization of a great variety of rocks results in a convergence to hornfels.

A number of contacts show a series of zones of differing material. The zones grade into each other with no sharp plane of separation (see pages 432 to 437 and the lists of further readings).

Regional Metamorphism.—Regional metamorphism is a more widespread process than contact metamorphism, either more distant from a magma than half a mile or at least more independent of the magma. Great masses many miles across may be simultaneously and similarly changed. Probably most of the action is deep seated in the earth's crust.

The dominant processes are recrystallization and granulation with locally some dehydration, metasomatism or other modification. There is commonly a convergence to some type of schist.

The factors that favor regional metamorphism are primarily stress and the heat of the depths of the earth. These contrast somewhat with the dominance of emanations in contact action, but the contrast is not so sharp as it sounds. The emanations into contact rocks are hot-water solutions, and most regional metamorphism involves hot-water solutions also. It is an open question whether the regional solutions are derived to any considerable extent from magmas at unknown depths below. Different cases of regional action with similar depth and pressure effects differ so widely in intensity of recrystallization, that they strongly indicate the effect of magmatic emanations in the more-intense alteration, even if no igneous rock is exposed near by (page 384). Batholiths may spread widely under the crust without exposures, and additions or metasomatic changes may be widespread around them. Most regional action, involving deep burial, probably has been distributed over prolonged periods of time.

The rocks most readily altered by regional metamorphism are the clays and associated sediments. Clay is quickly altered to slate and schist. Similar agents and forces acting on igneous rocks may have hardly any appreciable effect; others are intermediate. The mineralogic results of different compositions are discussed on pages 396 and 409.

Regionally metamorphosed rocks may show zones of varying intensity of alteration, but these seem to be related to depth or heat if not to the proximity of magma.

Hydrothermal Metamorphism.—Hydrothermal metamorphism is much less extensive than regional metamorphism and is localized by fractures in rocks rather than by magmatic invasion as in contact action. When hot solutions circulate through fractures, most rocks show a characteristic alteration. The walls of many mineral veins have been thus altered, and the prominence here given to this minor form of metamor-

phism is due to the fact that a great proportion of mine openings follow veins and encounter no other rocks than those hydrothermally altered. Few of the rocks have special names as metamorphic rocks (see page 372), but the alteration is none the less characteristic and in some places profound. (See the criteria tabulated for several depths.)

Hydrothermal alteration is almost wholly metasomatic and typically proceeds through successive stages of intensity, propylitization (Fig. 60), sericitization (Fig. 257), and silicification. Where the effects are intense, zones of rock farther from the fractures may show milder degrees of alteration. These changes are well shown by altered igneous rocks. In contrast to these, the first notable effect on a limestone may be a silicification.

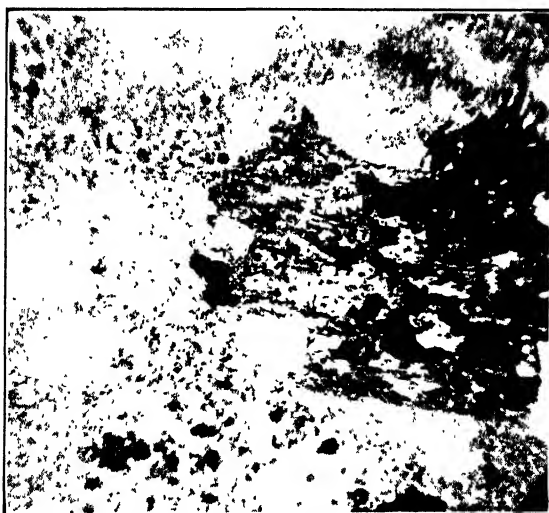


FIG. 257. —Hydrothermally altered (sericitized) quartz monzonite at Butte, Montana. Compare with the very slightly altered portion of the same rock shown in Fig. 60. Here the ferromagnesian minerals have disappeared leaving a pseudomorph (dark patch) consisting of quartz, sericite, and pyrite; feldspars have become patches of sericite; only the original quartz remains unaffected. Crossed nicols. $\times 70$.

The name hydrothermal indicates that hot-water solutions are the dominant factors in this type of alteration. The exact temperature of such solutions and the pressures needed to keep them from becoming gases are only roughly estimated. The temperatures of solutions that deposit ores are estimated as between 75° and 500°C . The nature of the materials in solution is noted on page 416.

The study of ores also gives us our best information as to the source of the solutions involved in hydrothermal action. Although both meteoric and magmatic waters may be present, the ores are so definitely zoned around some igneous masses, that the alteration is probably due in very large part to solutions of magmatic origin.

CRITERIA SUGGESTING HYDROTHERMAL ACTION AND ITS DEPTH

Usually found near granitoid rocks and porphyries, not usually deriving its hot water from lava flows.

A. Shallow hydrothermal alteration (epithermal).

Minerals. Most definite in italics. Stronger criteria if three or four are associated.

Gangue minerals: *adularia*, *alunite*, *zeolites*, *chalcedony*, *jarosite*, *nontremite*, *hydrargillite*, *quartz* as aureoles around phenocrysts, *analcite*, *stilbite*, *marcasite*, *carbonates*. (Fluorite, opal, barite, etc., may be cold-water minerals.)

Ore minerals: *cinnabar*, *stibnite*, *native copper*, *tellurides*, *selenides*, *silver*, *complex As-Sb-sulphosalts*. Chief ores are of *Au*, *Sb*, *Hg*, *Ag*, *Pb*, *Zn*.

Textures. Delicate colloform banding, crustified symmetrical banding, *pseudomorphs* of quartz after calcite in calcite cleavage planes, comb structure. *Open vugs*, *vesicles*, *amygdules*, etc., common. Short, irregular veins. Brecciation. Not usually slaty or schistose, or with metacrysts.

Suggestive characteristics:

1. Extensive silicification near the vein, especially by *chalcedony*.
2. Associated with rocks that form at shallow depths, and rocks that are weathered.
3. Ores are deposited in or very near fractures.
4. The usual epithermal sequence of formation is quartz, pyrite, sphalerite, chalcopryrite, galena, tetrahedrite, argentite, gold, electrum, sulphosalts.

B. Hydrothermal alteration at moderate depth (mesothermal).

Minerals. Most definite in italics. Stronger criteria if three or four are associated.

Gangue minerals: none that are positive criteria. Secondary orthoclase and albite are good signs.

Distinguished from hypothermal deposits by occasional *ankerite*, *barite*, *celestite*, *rhodochrosite*, *rhodonite* and *chert*.

Distinguished from epithermal deposits by occasional *apatite*, *orthoclase* and *siderite*.

	From
a. Pyrite.....	Magnetite
b. Sericite } Adularia }	Silicic feldspars Feldspathoids
c. Leucoxene } Rutile }	Ilmenite Titanite
d. Chlorite } Serpentine } Pyrite }	Mica Pyroxene Amphiboles Olivine
e. Epidote.....	Plagioclase Hornblende, etc.
f. Quartz and Carbonates }	Many minerals

Ore minerals: *chalcocite*, *enargite*, *argentite*, *orpiment*, *niccolite*, *penitlandite*, *petsite*, *realgar*, *bismuth*. Includes nearly all hypogene sulphides, arsenides, etc. Complex sulphosalts are particularly common. Chief metals: *Cu*, *Pb*, *Zn*. Coarse *bornite* is a very positive indication.

Textures. Finer grained than in deep seated zones; banded, crustified, *drusy* cavities; often fibrous; disseminated ores; pseudomorphs; fractured rocks.

Suggestive characteristics: Action most important at these depths.

1. Extensive widespread alteration, carbonation, sericitization, propylitization, silicification, related to fractures. Some dolomitization.
2. Presence of sericite, chlorite, calcite, epidote, serpentine, fibrous biotite, rutile, leucoxene, secondary orthoclase and albite.
3. Veins have smooth straight walls?
4. Andesites and basalts (less often rhyolite), are favorable; become bleached, basic rocks turning dull green, rhyolites chalky white.
5. All mafic minerals altered if attack is moderate

6. Absence of the characteristic products of weathering, unless the rock has been weathered later.
7. Zones from quartzose at the vein wall to sericite, and farther out to chlorite.
8. Occur near intrusive rocks, and near veins.
9. Not usually schistose.

C. Deep hydrothermal alteration (hypothermal).

Minerals. Most definite in italics. Stronger criteria if three or four are associated.

Gangue minerals: Garnet, tourmaline, scapolite, spinel, topaz, diopside, cummingtonite, cryolite, amphiboles and pyroxenes.

Ore minerals: Magnetite, specularite, pyrrhotite, cassiterite, molybdenite, scheelite, wolframite. Chief ores are Au, Sn, Fe, Cu, Mo, and W.

Textures. Intimate intergrowth of interlocking grains of ore and gangue similar to granitoid igneous rocks. Crude banding, never delicate or crustified, no comb structure. Coarse grain. Many fluid inclusions. Fissure walls apt to be irregular. Very low porosity.

Suggestive characteristics:

1. Very few complex salts; little native metal other than gold.
2. Alteration of wall rocks not extensive.
3. In rocks showing evidence of deep burial, such as schistosity, lack of porosity, etc. Mostly in old rocks.
4. Greisen, and luxullianite.
5. Rarely wide veins or large commercial deposits.
6. The relation to veins distinguishes from contact and segregated ores. Segregated ores rarely show primary carbonate, as many veins do; but might have olivine or feldspathoids, as deep veins do not.
7. The usual hypothermal and mesothermal sequence of formation is: Quartz, cassiterite, wolframite, pyrite, arsenopyrite, arsenides, stannite, sphalerite, enargite, tennantite, tetrahedrite, bornite, chalcopyrite, galena, chalcocite, sulphosalts.

SUMMARY OF COMMON METAMORPHIC CHANGES

Textural Changes.—The changes in size of grain differ with the dominant process of metamorphism. Recrystallization increases and granulation decreases the size of grains. Probably as a whole recrystallization dominates and grains become coarser, but the effect of granulation is by no means negligible.

The particular patterns or fabrics that result from metamorphism are listed on pages 352 to 356. Various intergrowths, differing in some respects from those formed in igneous and sedimentary rocks, are recognized, notably the hornfels texture and some poikilitic intergrowths. Probably most important is the recrystallization to a parallelism of needles or plates, a schistosity and cleavage.

Metamorphic banding is partly inherited from bedding in sediments and the banding of igneous rocks, but there may be some segregation into bands during the recrystallization. Much the largest part of the banding of metamorphic rocks, however, is probably due either to the elongation of lumps into sheets by rock flowage or to igneous injection along cleavage planes in schist.

The textures of granulation as distinct from those of recrystallization result from the breaking of grains into augen, mylonite and various small pieces; to some extent there are larger fractures and faults. Rocks become jointed and may develop shear zones of great extent. Granulated rocks also develop a cleavage, but it is different from that of recrystallized rocks.

Rock cleavage is the capacity of a rock to part along certain parallel planes more easily than along others. It may be original as in beds and flows, but it is secondary in most metamorphic rocks. Secondary cleavages are grouped as flow cleavage (slatiness, schistosity, and some gneissic structures) and fracture cleavage (slip cleavage, strain-slip cleavage, close-joints cleavage, fissility, and rift). Either kind of cleavage may be superposed on the other. Some rock cleavage is linear and some



FIG. 258.—Photomicrograph of andesite porphyry, showing part of a feldspar phenocryst that has a coordinate system of fracture cracks, partly healed with sericite. These cracks though fairly regular in direction are not in the directions of the feldspar cleavage. Plain light, $\times 100$.

tabular or platy. The most detailed studies of rock cleavage were made by Leith,¹ involving the examination of thousands of specimens and several hundred thin sections, also the benefit of criticism from many sources.

Flow cleavage is parallel to the greatest and mean or the greatest and least dimensions (or both or even the greatest alone) of the parallel constituent mineral grains. The excellence of this cleavage is roughly proportional to the degree of arrangement and to the inequality of the dimensions of the particles. The cleavages of the minerals, due to molecular arrangement, must be sharply distinguished from rock cleavage. Mineral cleavages are important aids to the parallel arrange-

¹ LEITH, C. K., Rock cleavage, U. S. Geol. Survey Bull. 239, 1905.

ment of grain in producing rock cleavage, but they rarely if ever cause a cleavage in any other direction than that caused by dimensional parallelism. Minerals like mica and hornblende show both dimensional and crystallographic parallelism, whereas quartz is seldom oriented crystallographically.

Fracture cleavage is independent of the grain but is conditioned by closely spaced fractures having a fairly uniform direction, some of which may be minute, or incipient, or partly healed or healed with soft minerals (Fig. 258). Even igneous rocks have a rift or direction of easy cleavage (pages 39 and 59).

The relations of cleavage directions to forces applied have been discussed at length.¹ Cleavage may be normal to the major pressure or may be on shearing planes inclined to the stress. Although the first is commonly a flow cleavage and the second commonly a fracture cleavage, there are many cleaved rocks in which it is hard to distinguish whether the elongation was due to flow or shear.

Common Changes in Rocks and Minerals by Metamorphism.—The changes that take place are so different in different kinds of metamorphism, that the attempt here made is simply to select the most common. It seems to be generally agreed that most metamorphic rocks show signs of recrystallization and dynamic action. Such action turns most rocks into schists and gneisses. The contact rocks—hornfels and silicate-carbonate zones—are not to be listed as common, no matter how important commercially some may be.

TABLE XXIII.—SUMMARY OF ABUNDANT RESULTS OF METAMORPHISM

Original rock	Common slight metamorphism	Extreme metamorphism
Intermediate and basic igneous rocks: Basalt, diabase and andesite.....	Propylite and chlorite schist	Hornblende schist
Pyroxenite and peridotite.....	Hornblende schist and serpentine	Chlorite schist
Diorite and related rocks.....	Diorite gneiss	Mica schist
Silicic igneous and arenaceous rocks: Granite, rhyolite.....	Granite gneiss	Sericite schist Mica schist
Sandstone, arkose.....	Quartzite, arkosite	Granite gneiss
Argillaceous rocks: Clay, shale, etc.....	Slate, phyllite, and mica schist	Granite gneiss
Calcareous rocks: Limestone, dolomite.....	Marble	Serpentine Soapstone

¹ LOVERING, T. S., The fracturing of incompetent beds, Jour. Geology, vol. 36, pp. 709-717, 1928.

Even with this restriction of the common processes to recrystallization and dynamic action, results may differ depending on which of the two dominates and depending also on how intense the action has been. The table on page 429 indicates two degrees of intensity of effects on different kinds of rocks but shows nothing as to whether dynamic action or recrystallization is dominant. Probably as a rough generalization dynamic action is prominent in the change of massive igneous rocks to schists, whereas recrystallization dominates hydrothermal changes in igneous rocks and most of the changes in sediments. The most useful general conclusions from the table are that biotite schists are more commonly of sedimentary than of igneous origin and that the gneisses and the hornblende and chlorite schists are largely of igneous origin.

The mineral changes tabulated are abstracted from tables in the "Treatise on Metamorphism" by Van Hise.

TABLE XXIV.—COMMON MINERAL CHANGES IN METAMORPHISM

Mineral	Source
Actinolite.....	Pyroxene and olivine
Albite.....	Plagioclase and zeolite
Biotite.....	Clays, pyroxene, amphiboles, magnetite and feldspars
Calcite.....	Pyroxene, amphibole, epidote, garnet, etc.
Chalcedony....	Many silicates
Chlorite.....	Garnet, biotite, pyroxene, prehnite, staurolite, tourmaline, vesuvianite, clay
Cordierite.....	Clays
Epidote....	Feldspars, pyroxene, biotite, amphibole
Garnet.....	Clays, chlorite, limestone, etc.
Hematite and magnetite.....	Many iron silicates
Hornblende.....	Augite, chlorite, garnet
Muscovite, sericite, paragonite.....	Feldspars, feldspathoids, clays, etc.
Orthoclase....	Feldspathoids, zeolites, clays
Prehnite.....	Feldspars and zeolites
Quartz.....	Many silicates
Serpentine.....	Olivine, pyroxene, micas, amphiboles, garnet
Sillimanite....	Clays, andalusite, kyanite, mica
Staurolite....	Clays
Talc.....	Olivine, pyroxene, amphibole, micas, garnet, etc.
Titanite.....	Ilmenite and rutile
Tremolite.....	Pyroxene, olivine, calcite and dolomite
Wollastonite....	Calcite and dolomite
Zoisite.....	Plagioclase, garnets, etc.

TABLE XXV.—EXPECTED ALTERATION PRODUCTS FROM COMMON MINERALS

Mineral	Products
Augite.....	Hornblende, biotite, chlorite, quartz, epidote, magnetite
Biotite.....	Chlorite, epidote, magnetite, pyroxene, muscovite, quartz, etc.
Calcite.....	Dolomite, wollastonite, garnet, biotite
Chlorite.....	Garnet, biotite
Clays.....	Biotite, muscovite, staurolite, garnet, feldspar
Garnet.....	Chlorite, epidote, hornblende, pyroxene, calcite, quartz, magnetite, serpentine
Hornblende....	Augite, biotite, chlorite, epidote, magnetite, calcite, quartz, serpentine
Ilmenite.....	Leucoxene, rutile, titanite
Magnetite.....	Pyrite, biotite, etc.
Muscovite.....	Paragonite

TABLE XXV.—EXPECTED ALTERATION PRODUCTS FROM COMMON MINERALS.—(Cont'd.)

Mineral	Products
Nephelite.....	Albite, zeolites, muscovite
Olivine.....	Serpentine, iddingsite, biotite, amphiboles, pyroxenes, magnesite, quartz, magnetite
Orthoclase and microcline....	Muscovite, sericite, epidote, calcite, quartz
Plagioclase.....	Zeolites, paragonite, epidote, prehnite, zoisite, quartz, albite
Prehnite.....	Chlorite, quartz
Serpentine.....	Magnesite, opal, quartz, hematite
Staurolite.....	Chlorite, quartz
Tourmaline....	Biotite, chlorite
Zoisite.....	Calcite, quartz

ORIGIN OF A METAMORPHIC ROCK

In a large majority of petrographic problems the question is not what to expect in the metamorphism of a particular rock but rather what to think the metamorphic rock was before it was metamorphosed. This requires all the petrographic skill of the most experienced men, and the criteria are often revised and much discussed. The tabulation here given is, of course, tentative and must be revised as knowledge of such rocks improves. The criteria for determining in metamorphic folded rocks (1) the top and bottom of beds and (2) the presence of an unconformity were cited on page 325.

CRITERIA SUGGESTING THE ORIGIN OF METAMORPHIC ROCKS

Some of the field relations are the only ones that can be applied with even a moderate degree of accuracy; the others should be applied with care and inferences should be drawn only when a number of them check. (See the list of further readings.)

Structural and textural relations.

Igneous.—Gradation into igneous rocks. Preservation of textures in the less altered portions, such as porphyritic, graphic, diabasic, amygdaloidal, ellipsoidal, chilled contacts. Preservation of structures, such as the outline of a dike. Uniformity over large areas (?). Irregularity of form of distribution (?). Augen, mortar, and mylonite are likely.

Sedimentary.—Gradation into sedimentary rocks. Preservation of sedimentary structures and features, such as bedding, stratification, fossils, conglomerate and pebbles, fragmental relics, oolitic, ripple marking, secondary growth. Banded structure, with persistency in composition throughout the band. Abrupt change in texture and composition across such bands within very narrow limits. Intercalation of beds of limestone or quartzite. Rusty weathering (?). Recrystallization more common than augen or granulation textures.

Mineral composition.

Igneous.—Preponderance of feldspar. Plagioclase limited to one variety. Nephelite, allanite. Presence of the accessories zircon, monazite, titanite, xenotime, garnet, etc., if not abraded. The gneisses are mostly igneous. Chlorite, epidote, and hornblende schists are often igneous. Rock variation is from mafic to felsic.

Sedimentary.—Preponderance of quartz; preponderance of calcite; a variety of plagioclases. Abundant development of aluminum silicate minerals, staurolite or axinite or cordierite; wollastonite; less definitely garnet. If talc is dominant, it is usually derived from a dolomite. Graphite in considerable amount and evenly distributed. Absence of accessories in argillaceous rocks, or if present much abraded; rutile needles. The quartzites and marbles are sedimentary. Most of the slates, phyllites and biotite schists are sedimentary. Rock variation is from siliceous to aluminous to calcareous.

Chemical composition.

Igneous.—Negative of all points under sedimentary.

Sedimentary.—Dominance of magnesia over lime, potash over soda and ferric over ferrous iron—especially if together.

Molecular ratio of alumina to the sum of soda, potash, and lime is greater than 1. Silica in greater abundance than in most igneous rocks.

Variations from the composition of igneous rocks as shown by tables of analyses.

NOTE.—The igneous tufts and breccias are transitional from igneous to sedimentary rocks and when altered are especially doubtful. They have usually only fine-grained or glassy fragments, usually only igneous fragments and are associated with flows.

NOTE.—Many injection gneisses are *both* igneous and sedimentary in origin. The igneous portion so strengthens most such rocks that they resist further deformation, and the injected bands are rarely metamorphosed.

DEGREES OF METAMORPHISM AND ZONES

A considerable literature has grown up around the idea that it may be possible to determine the depths and the temperatures of metamorphism by the results. See further readings. Up to date the several kinds of effects are so confused in the discussion that no generally acceptable scheme has been evolved as to nomenclature and classification. Nevertheless the data for particular districts are of great interest and some generalizations may yet be possible.

Different petrologists have emphasized different features as signs of profound metamorphism, and the criteria used are here collected. Many have used only the mineralogical features and based the whole discussion on groups of minerals. It should be clear, however, to the most casual observer that there may be different degrees of intensity of effects produced by heat and pressure, quite aside from the several degrees of intensity of effects produced by water or by magmatic emanations. These factors combine to produce an almost infinite variety of effects, which can not be compiled into one single series.

It should also be clear that the results of metamorphism depend not only on the intensity of the process but also somewhat on the nature of the original material. A gneiss is formed from granite by much less intense change than is needed to make gneiss from a clay.

The chief points ordinarily noted as signs of varying intensity are the minerals formed by recrystallization; some form earlier than others. The sequence may be a somewhat useful guide, as most metamorphic rocks show recrystallization under conditions of pressure, heat and water. The rarer cases of dry heat and of pressure in the absence of heat result in a relatively small proportion of our metamorphic rocks. On this basis we can refer to the sequence of formation of minerals in recrystallization (page 409) and the sequence in the alignment of minerals (page 357) as indications for judging the intensity of metamorphism.

Beyond that we have the occurrence of zones around intrusives, but there is no great regularity in such zones.

CRITERIA OF INTENSITY OF METAMORPHISM

The *degree of metamorphism* refers to a rough measure of the *extent of change* in a rock. It does not refer to the degree of heat or degree of pressure alone; a granite formed at high temperature and pressure is not much metamorphosed by high temperature and pressure. All criteria must be based largely on a knowledge of the original rock.

Change in composition.

How far the minerals (pages 434 to 437) differ from the minerals of the original.

Surface luster or gloss.

Proportion of recrystallized minerals.

New textures or modes of aggregation.

Degree of granulation.

Deformation of original structures and textures.

Proportion of metacrysts.

Elongation of grains.

Parallelism of grains.

Change of size of grain.

Rock cleavage independent of original structures.

Change of density.

Rocks produced (page 429).

Low intensity in a shale leaves fine grain.

Low intensity in a coarse rock leaves only strain shadows.

Metamorphic Zones around Igneous Intrusives.—The most accurate observations of the different effects produced by varying intensity of metamorphism are made at intrusive contacts. Outside the igneous rock (which may have endomorphic effects) there are in many places a series of zones, and it is perfectly clear that the greater heat and more abundant emanations near the magma produced the greater effects. The nature of the contact rock largely determines the probability of a zoning of the effects. Sands become quartzite, and limestone becomes marble or hornfels depending on the purity of the lime, but such rocks show little zoning except in the degree of cementation or recrystallization of the original minerals. The zones in clay rocks are more complex and interesting and show some similarity in the several cases studied.

In argillaceous rocks contact action usually finds the country rock of the outer zone a slate of fine grain with a specific gravity about 2.75. Nearer the intrusive are mica schists with garnet and andalusite and a gravity of 3.18. Then comes a zone of staurolitic and sillimanite schist, gravity 3.23. The inner zone is more gneissic, with hornblende or augite, feldspar, kyanite and often with minerals showing magmatic additions or emanations. Its gravity may be as great as 3.6.

The nature of the zones varies not only with the nature of the sediment and the distance but also to some extent with the type of intrusive magma, gabbros normally producing a less hydrous contact zone than granites.¹

The zonal arrangement of ore deposits around intrusives² is also suggestive of a zoning of metamorphic action. As some of the ore zones

¹ SCHWARTZ, G. M., The effect of granite and gabbro intrusions on the Ely greenstone, Jour. Geology, vol. 32, pp. 89-138, 1924.

² EMMONS, W. H., Primary downward changes in ore deposits, Am. Inst. Min. and Met. Eng. Trans., vol. 70, pp. 964-992, 1924.

may extend several miles from the outcrops, magmatic emanations no doubt contributed some material to the processes of regional metamorphism to an equal distance.

Zones Related to Depth.—The following tables, modified from the studies of Grubenmann and Niggli, show the supposed relation of rocks and minerals to the zone of formation. Figure 259 may help to fix in mind some of the most common features of these zones. As noted on page 432, however, the results vary with the original. Although the epi-zone is near the surface and the kata-zone most remote, the three zones are not sharply marked off from each other and not strictly related to depth. They are zones of conditions, and especially at times of magmatic invasion these zones move up more or less toward the surface.

The general effects of the upper zones (below weathering) are hydration and carbonation; those of the deep zone are silication, dehydration, deoxidation and decarbonation.

TABLE XXVI.—METAMORPHIC ROCKS BY ZONES OF FORMATION

Epi-zone	{ Characterized by phyllites of many kinds. Sericite schist, chlorite schist, talc schist, serpentine, quartzite, crystalline limestone and various sheared phases of massive rocks
Meso-zone	{ Characterized by schists of many kinds. Amphibolite, granulite, mica gneiss, amphibole gneiss, garnet gneiss, quartzite, marble, skarn, greisen
Kata-zone	{ Characterized by gneisses of many kinds. Biotite schist, garnet rock, wollastonite rock, hornfels, skarn, granulite, eclogite, augite rock, marble and quartzite, luxullianite, leptite

On the basis of finding such facies zones of metamorphism have been mapped in a few places where they are not visibly related to igneous intrusion. (See the list of further readings.)

Zones of Hydrothermal Effects.—As noted on page 425, the wall rocks of many veins show a series of effects that are most intense at the immediate contact. The walls of quartz veins may be silicified, whereas the rock a few feet away is sericitized and a wider zone beyond has only a little sericite with more chlorite.

CYCLES OF METAMORPHISM

General Features.—The processes of metamorphism are so varied and complex that a sequence of processes is commonly found to have affected the formation. If this sequence is somewhat orderly, it may return to its starting point and begin as a new cycle of changes. Perhaps more often the sequence is not so regular as to constitute a cycle, but certain cycles are common enough to be noteworthy.

Leith and Mead¹ (using the term metamorphism in the broad sense to include weathering) find a cycle of effects somewhat analogous to the physiologic cycle of metabolism (Fig. 260). The anamorphic part of the

¹*Op. cit.*, Plate I,

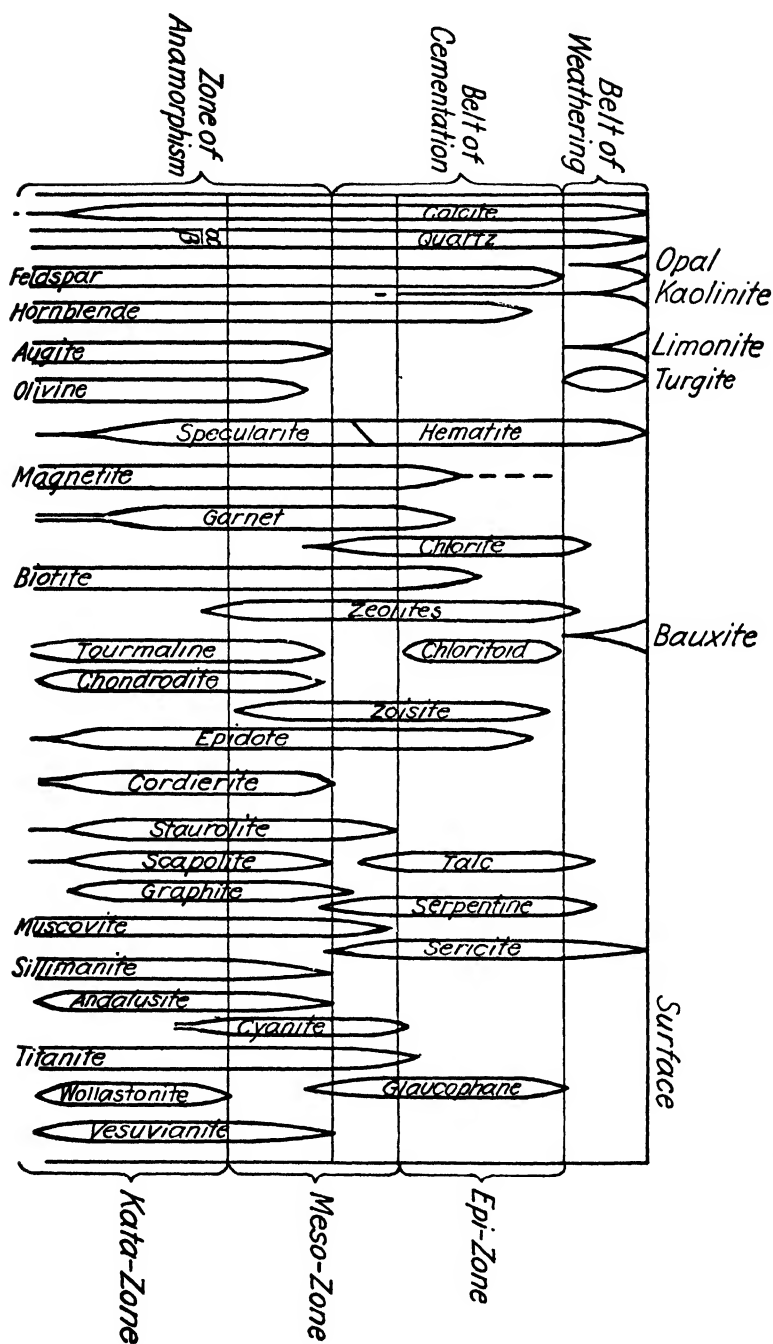


FIG. 259.—Diagrams of the zones of formation and stability of the common minerals. Those that extend to the bottom of the diagram may also appear in igneous rocks.

TABLE XXVII.—MAIN MINERAL FEATURES OF THE METAMORPHIC ZONES
(After Grubenmann and Niggli)

	Si	Ca	Mg	Si + Ca	Si + Mg, Fe	Mg, Fe + Al ± Si	Si + Ca + Al ± Mg, Fe	Si + Al(Fe) + Na	Si + Al + K	Si + Al + K + Fe + Mg	Si + Al	Fe	Ti
Epi-zone	Quartz	Dolomite Calcite	Daweylite Magnesite	Quartz Calcite	Serpentine Talc Stilpnomelane	Chloritoid Chlorite	Epidote Zoisite	Albite	Sericite	Sericite + chlorite	Sericite	Siderite Pyrite Hematite Magnetite	Titanite Rutile
			Hydro- magnesite Bucrite		Actinolite	Almandine		Paragonite Glaucoophane	Muscovite				
Meso-zone	Quartz	Calcite	Bucrite		Actinolite Hornblende Cummingtonite Anthophyllite Grunerite	Biotite Phlogopite Almandine (Pyrope) Staurolite	Hornblende Epidote Zoisite	Albite Oligoclase		Biotite		Hematite Pyrite Magnetite	Titanite Rutile
								Sodic hornblende	Microcline				
Kata-zone	Quartz	Calcite	Periclase	Woll- astonite	Orthorhombic pyroxene	Biotite	Labradorite Bytownite Grossularite Vesuvianite Scapolite	Plagioclase Jadite			Andalusite Sillimanite	Magnetite Pyrrhotite	Titanite Rutile Ilmenite Perovskite
					Hornblende Olivine Hornblende group Diopside Augite Monticellite	Cordierite Garnet Spinel Hedenbergite Andradite		Sodic hornblende Sodic augite	Orthoclase	Biotite			

cycle rarely goes to the complete regeneration of magma but gives a crystalline rock which may again start through the cycle.

The changes in such a cycle are controlled by conditions, and the normal sequence of metamorphic conditions may be worthy of more

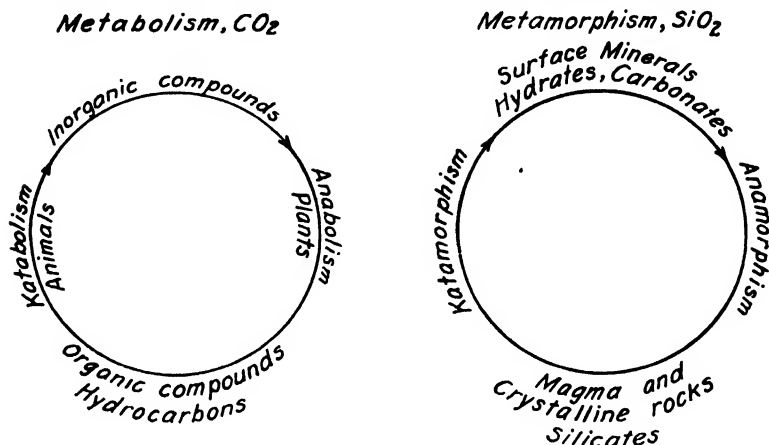


FIG. 260.—The cycle of metamorphism compared with the cycle of metabolism.

detailed notes, even if we exclude that portion of the cycle usually classed as weathering.

Normal Sequences of Geologic Events in Metamorphism. *The Dynamic Sequence.*—The sequence of simple dynamic metamorphism may

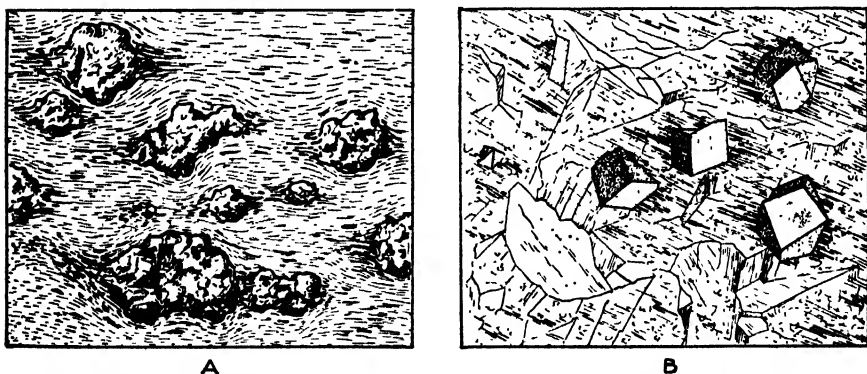


FIG. 261.—Metacrysts of garnet in chlorite schist. A. Metacrysts formed before the maximum of metamorphic deformation are granulated to augen. B. Metacrysts formed after the deformation produced a schistosity sharply transgress the foliation. About natural size.

be illustrated by the changes in sediments. The most abundant sediment, clay, is buried under a load of other sediments and surface rocks until the pressure at the base of a geosyncline of such material is enormous. The conditions of load produce a certain degree of static metamorphism,

but without dynamic or igneous action the changes are chiefly a recrystallization to argillite, possibly with some metacrysts.

At the climax of orogeny the geosyncline is crumpled, and the deformation produces schists and possibly gneisses, especially if igneous injection accompanies the deformation. This is the period of most prominent dynamic metamorphism. Early structures are largely obliterated.

During this climax of deformation the differential stresses are relieved, but the load persists and may even be largely increased, because of the thickening of overlying sediments by folding or by overthrust. Under this great load, whether increased or not, static metamorphism continues, and again metacrysts may develop. Before the rocks can be exposed to view the great load must be eroded, but no normal changes during the removal of the load obliterate the late metacrysts until they reach the belt of surface weathering¹ (see Fig. 261).

The time involved in burial and erosion may be long compared with the time of active deformation, but the changes in texture and structure are relatively great in the shorter dynamic period. This maximum of effects, however, is sufficiently prolonged to allow deformation to occur by rock flowage and recrystallization.

Deformation and metamorphism may locally be important after the maximum and during the progress of erosion. The changes that occur at more moderate temperatures and pressures after the maximum are katamorphic as the term is used by Leith; "retrogressive" if one considers that the most intense heat and pressure produce the most intense metamorphism. At least the minerals of the deep zone may be replaced by others characteristic of the shallow zone. There are commonly pseudomorphs, such as that of chlorite after garnet and of sericite after andalusite.

At any particular stage in the sequence just outlined, there may come a change of conditions such that the normal sequence is interrupted. Even if the sequence is complete and the altered sediment is exposed at the surface, it may again form sediment or be covered with new sediment and start the sequence again. The processes are therefore properly called a cycle.

The Contact-metamorphic Sequence.—The sequence of events during contact metamorphism may also give a cycle, for a region of intrusive igneous rocks may be found to have been intruded at more than one period. Probably in general the contact cycle is: (a) recrystallization by heat, (b) pervasive emanation and reaction, (c) hydrothermal introduction and attack. In alteration of limestone it is not rare to find that after the garnets of the second stage are formed, there are later veins and replacements of the garnet by carbonate. Such a carbonated rock would

¹Leith, in "Structural Geology," p. 185, gives six reasons for believing that metacrysts are late to form. Those we see are no doubt the late ones, for early metacrysts must be largely destroyed in the flowage at the climax of orogeny.

be a not unfavorable host for later cycles of metamorphism. Some contact rocks may also have been complicated by early emanations in addition to late hydrothermal attack.

The Sequence during Batholithic Invasion.—There is a rather complex sequence, also, in batholithic invasion and metamorphism, and several batholiths have intruded older batholiths, giving a repeated cycle of effects. The sequence is so different in different batholiths, that it is difficult to select a particular sequence as standard for the cycle. Some seem to accompany folds (page 200) and regional metamorphism, but others form after the folding and without much metamorphism.

The batholiths with folded and metamorphosed walls were probably deeper in the crust than others when invaded and show a fairly character-

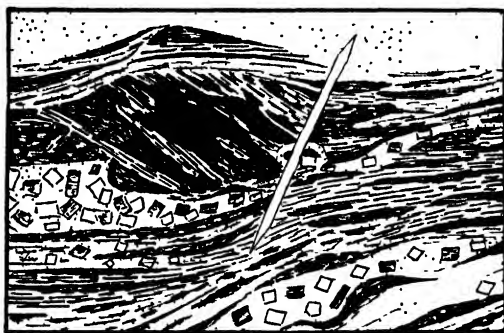


FIG. 262.—Sketch of a polished slab of gneiss from Morton, Minnesota, about 6 feet wide and 4 feet high. An inclusion of hornblende (black) has reacted with and been partly assimilated by the granite (stippled) which has normally a pink color. The granite magma has soaked into the hornblende rock making it a gray gneiss (lined areas). The lighter bands in the gneiss are still granitic in composition but are gray from reaction with the hornblende. Locally the granite has coarse feldspar (blocks) approaching a pegmatite, and this phase crosses the gray gneiss. The latest introduction crossing all the earlier phases is a gash vein of quartz.

istic sequence. First, the magma injects its schistose walls *lit-par-lit*. Probably at the same time the walls and border zone are deformed by the upward and outward “bursting pressure” of the magma column. This altered material is seen as metamorphic rock only rarely, because, in the first place, it is so deep that only profound erosion can expose it, and, in the second place, it is an early effect, which is likely to be obliterated by the advance of the main magma. The early *lit-par-lit* and quartz veins constitute an advance guard, but their metamorphic effects may be stopped in or dissolved later. When the main batholith magma is about ready to cease stopping and assimilating, its effects, though produced with much less violence than the earlier ones, are in a position to be preserved until the rocks are cooled and eroded. There may be some injection, some contact metamorphism, some stopping of injected blocks, all of which remain near the contact. As the magma crystallizes to form a relatively

anhydrous rock, the hydrous mother liquor may inject and alter the wall rocks already altered by early emanations (Fig. 262). This late injection results usually from some earth movement causing mashing or filter pressing. In fact, if the batholith is large, there is time for a series of alternate injections and mashings, only the latest pegmatites in some districts being free from dynamic effects.¹

In summary, the cycle recorded in many cases of batholithic injection includes a peripheral schistosity of the walls, an early injection and breaking or stoping of the schist, a late injection of pegmatite and an alternation of squeezing and injection or filter pressing, which finally makes the intrusive itself largely gneissic, especially in those phases earliest to solidify.

CRITERIA OF GENERAL SIGNIFICANCE

Two sets of petrographic criteria are not to be logically considered under the heading of any particular class of rocks or processes but only after all have been discussed. These are the criteria of the temperatures to which a rock has been subjected either in depth or near igneous intrusives—the geologic thermometer—and the criteria of the order of formation and crystallization of minerals.

CRITERIA SUGGESTING THE TEMPERATURE OF FORMATION

(The Geologic Thermometer)

These are largely from Bowen² and are based on melting points, transformations, dissociations, breakdown of solid solutions, and observed lavas.

Igneous rocks.

Igneous rocks soften between 960° and 1160°C. As a rule the more silicic rocks solidify at 800° to 700°C.; pegmatites below 600°C.

Alpha quartz (long prisms), below 573°C.

Beta quartz (short prisms), 573° to 870°C.

Tridymite, 870° to 1470°C.

Cristobalite, 1470° to 1720°C.

Native iron with Widmannstätten structure, 1100° to 1400°C.

Olivine, 1175°C. ±, forsterite below 1890°C.

Mica forms only below 800°C., biotite dissociates at 850°C.

Diamond forms above 690°C.

Hornblende below 550°C., augite and brown hornblende above 850°C., enstatite below 1557°C.

Cancrinite, above 1258°C.

Plagioclase, below 1550°C.

Contact-metamorphic rocks 300° ± to 800°C.

Pyrrhotite, above 550°C. Pyrite changes to pyrrhotite 600° to 800°C.

Birefringent garnet, below 800°C.

Quartz, alpha, below 573°C., beta, above 573°C.

Wollastonite, below 1180°C., pseudowollastonite above, is not known in nature.

Tremolite, 700° to 1100°C.

Kyanite above, sillimanite below 1320° to 1380°C.

Development of Ca, Mg, and Fe silicates from limestone and dolomite begins at about 500°C.

¹ BARRELL, JOSEPH, Igneous invasion and regional metamorphism, *Am. Jour. Sci.*, 5th ser., vol. 1, pp. 263–267, 1921.

² BOWEN, N. L., Geologic thermometry in Fairbanks *et al.*, "The Laboratory Investigation of Ores," pp. 172–199, 1928.

Tourmaline, up to 800°C., but small crystals may form below 300°C.

Topaz, higher temperatures than tourmaline.

Specularite, above 358°C.

Natural coke, 400° to 500°C.

Calcite dissociates at 900° to 1100°C. at 20 atmospheres and higher under more pressure.

Hydrothermal deposits 100° to 500°C.

Pyrrhotite, above 550°C., below 1187°C.

Alpha quartz, below 573°C.

Aragonite, below 450°C.

Marcasite, below 450°C.

Albite, above 400°C.

Orthoclase, above 340°C.

Adularia and zeolites, 100° to 300°C.

Silication, above 260°C.

Jasperoid, 200° to 300°C.

Gypsum, below 30°C.; anhydrite, above 66°C.; either between 30° and 66°C.

Sulphur crystallizes below 96°C.

Quartz with liquid and gas inclusions.

(See Figs 33 and 263.) Use with caution for quartz in any form of occurrence.

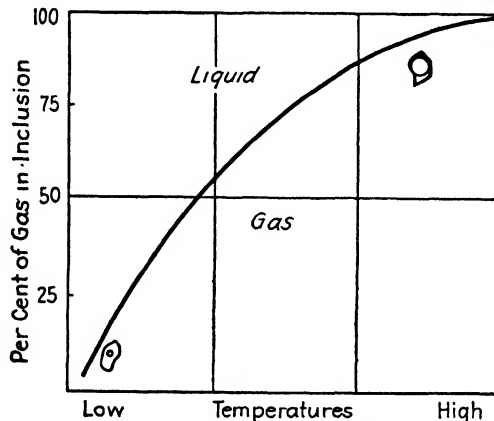


FIG. 263.—Curve of the approximate relative sizes of gas bubbles in liquid inclusions in quartz; to be used with caution, and by comparison with known material rather than by giving definite values to the temperatures. See Holden, *Am. Mineralogist*, vol. 10, p. 219; and Bowen, in Fairbanks, "Laboratory Investigation of Ores," pp. 192–194.

CRITERIA SUGGESTING THE ORDER OF FORMATION OF MINERALS IN ROCKS

A. In igneous rocks.—(See page 44.)

1. A mineral enclosed in another is earlier.(?)

2. Phenocrysts are earlier than the groundmass.

3. Fringes, hydrothermal and deuteric effects usually late.

4. Simultaneous deposition gives mutual anhedral and at times graphic intergrowths.

5. Borders at the top and bottom of a mass are earlier than the center.

6. Expected order is: (1) ores, (2) ferromagnesian minerals, (3) feldspars and feldspathic minerals, (4) Quartz.

Order in igneous ores; apatite, magnetite, hematite, pyrrhotite, pentlandite, chalcocite, bornite, zircon, and titanite.

B. In sedimentary rocks.

1. The fragmental grains are older than the cement. (Such as secondary growth.)
2. The bottom of a bed is earlier than the top. (Criteria for top of bed.)
3. Minerals unstable under conditions of sedimentation are probably later.

C. In contact rocks and replacements. (See criteria of contact action.)

1. A mineral that cuts across recognizable structures is later than the minerals of the structures. For example, crystals projecting into sand grains, crystals enclosing some schistose grains, crystals intersecting bedding planes, and crystals intersecting fossils.

2. Pseudomorph is later than original.
3. Minerals following contacts or cleavages of other minerals are later.
4. Ice cake structures are remnants of earlier minerals.
5. The expected order in contact rocks is: pyroxene, epidote, garnet, magnetite, calcite, pyrite, pyrrhotite.

D. In regional metamorphic rocks.

1. Metacrysts are later than the matrix. (See Leith and Mead for sequence of metacrysts.)
2. Crystals showing relatively great strain or fracturing are early.
3. Fresh crystals are usually recrystallized and late.

E. In all rocks.

1. Hydrothermal action.
 - a. The expected order is: chlorite, etc., sericite, carbonates and pyrite; quartz.
2. Incongruity of mineral associations.
 - a. Calcite in basalt; basalt must have been present first.
 - b. Depends upon indirect geological or geochemical evidence.
 - (1) Garnet in a limestone.
 - (2) Pyrite in a sandstone.
3. Fillings of cavities are later than the walls.

PART VIII

MINERAL TABLES AND READINGS

TABLES OF ROCK-FORMING MINERALS

The following tables for different methods of identifying minerals are rather differently constructed. Some rare rock minerals are omitted, and if the student has reason to expect rare minerals, larger tables should be consulted (see page 9 and the readings on methods). The rare minerals of pegmatites are listed only in the table for the use of index oils, because thin sections of pegmatite are rarely satisfactory, and the oils furnish a very convenient method of identification. A list of the tables follows.

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In using Table VII, the specialist in sedimentary rocks should refer first to the column showing shapes of grains. On the other hand, a concentrate from igneous rocks, or one from an unknown problem is probably best tested for index of refraction before other work is begun.

The following abbreviations are used in the tables. Those minerals marked with an asterisk are most common in another main division of the table.

v. = violet. b. = blue. g. = green. y. = yellow. o. = orange.

r. = red. blk. = black. br. = brown. gr. = gray.

Main color in capital letters.

Str. = strong. Sl. = slight. Var. = variable.

Ac. = acicular. Pl. = platy. Rad. = radiating. Tab. = tabular.

Fib. = fibrous. Pr. = prismatic. Rect. = rectangular. Tetr. = tetragonal.

Hex. = hexagonal. Pyr. = pyramidal. Rh. = rhombic. Tri. = triangular.

Ir. = irregular.

X = fast ray vibration direction.

Y = medium ray vibration direction.

Z = slow ray vibration direction.

c = vertical crystallographic axis.

Ext. = extinction.

Sym. = symmetrical.

Interf. = interference.

Uni. = uniaxial.

Perf. = perfect as to cleavage.

Pleoch. = pleochroism.

Incl. = inclusions.

Cl. = cleavage.

G = specific gravity.

H = hardness.

Sol. = soluble rather easily.

Elong. = sign of elongation.

Bx. = bisectrix.

The names of common rock minerals are in the most prominent type, fairly common minerals in medium type, and rare ones in small type.

TABLE I A.—ISOTROPIC MINERALS
In Oils

Index	Shape	Remarks	Name
Very low.....	Rounded, etc.	Inclusions, color fringes	Gas and liquids
1.43.....	Triangular	{ Polygons, 3, 4, 5, 6 sides, 60°, 120° }	FLUORITE
1.44 ± 3.....	Irregular	{ Colorless, blue, yellow, etc. }	OPAL
1.456.....	Irregular	Always colorless, cavity filling	alum
1.47 - 1.55.....	Irregular	Colorless, earthy, no cleavage	Halloysite
1.48 - 1.52.....	Irregular	Colorless, yellow, forms matrix for crystals	SILICIC GLASS
1.48.....	Irregular	Colorless, blue, green, brown, 6 poor cleavages, large crystals	SODALITE
1.487.....	Irregular	Always colorless, basic rocks and cavities	ANALCITE
1.495 ± 5.....	Irregular	Poor cleavages, in igneous rocks, contain inclusions	{ Hauynite
1.50 ±	Irregular	Blue, in contact rocks	Noselite
1.508.....	Irregular	Gray or colorless, multiple twin inclusions, zoned	Lazurite
1.525.....	Irregular	In granites	LEUCITE
1.53 ± 1.....	Irregular	Groundmass for crystals	pollucite
1.53 -	Irregular	Amorphous or cryptocrystalline	ANDESITE GLASS
1.535.....	Platy	Colorless or blue. Soluble in water	CHALCEDONY*
1.544.....	Cubic	Brown, forms matrix for crystals	apophyllite*
1.54 - 1.63.....	Irregular	Brown, veins and altered mafic minerals	HALITE
1.55 ± 4.....	Irregular	Green, in iron formation	Basic glass
1.60 ± 3.....	Platy	Green or brown, confused aggregates	hisingerite
1.64 ± 5.....	Platy	Brown, <i>always colored</i>	Greenalite
1.69.....	Irregular	Brown	BIOTITE*
			ALLANITE*

* Usually anisotropic but may appear isotropic.

TABLE I A.—ISOTROPIC MINERALS.—(Continued)

Index	Shape	Remarks	Name
1.73 ±	Cubic	White, etc. Contact rocks	{ periclase helvite danalite Spinel GARNET
1.738.....	Irregular	Yellowish, in pegmatite	
1.74 ± 2	Irregular	Green, red, brown, blue, basic rocks	
1.80 ±.....	Irregular	Pink, brown, colorless	
1.8—	Irregular	Always colored, brown, red, black, almost opaque	CHROMITE
1.812.....	Cubic	Yellow, in alkalic rocks	
1.98 ± 2	Irregular	Brown, in igneous rocks	
2.20.....	Irregular	Brown, nearly opaque, in pegmatites	
2.23.....	Irregular	Red brown, in pegmatites	
2.37.....	Triangular or irregular	Brown, green, yellow, white, 6 cleavages	
2.38.....	Cubic or irregular	Red, brown to white	Sphalerite
2.42.....	Octahedral	Colorless, in gravel, or peridotite	
High.	Irregular	Brown to black, nearly opaque	
			Perovskite
			Diamond
			uraninite

TABLE I B.—ANISOTROPIC COLORLESS MINERALS WITH LOW INTERFERENCE COLORS
Birefringence Up to 0.020
In Oils

Alpha	Biref.	Optic axes	Optic sign	Ext. angle	Shape	Remarks	Mineral
1.36 ±	0.005	2	+	44°	Ir. Rect.	H = 3. Ores associated	cryolite
1.470	0.004	2	+	Ir.	Pseudo-hexagonal	Tridymite
1.478	0.012	2	+	9°	Needles	In amygdulæ, etc.	Natrolite
1.485	0.003	1	+	Ir.	May be biaxial	Cristobalite
1.486	0.001	2	±	In cavities	Analcite
1.494	0.005 ± 2	2	+	0°-10°	Tab. or Rad.	In cavities	Stilbite
1.498	0.007	2	+	Globular	Rarely fibrous	heulandite
1.504	0.012	2	+	2°-8°	Pl.	petalite
1.508	0.003 ± 1	2	+	Pseudo-isometric	LEUCITE
1.519	0.007	2	-	0°-5°	Round Ir.	Carlsbad or no twins	ORTHOCLASE
1.519	0.007	2	-	0°-5°	Ir. Pl. Pr.	Gridiron twins	MICROCLINE
1.521	0.009	2	+	0°-45°	Pr. Pl. Pr.	H = 2; dispersion	GYPSUM
1.525	0.008 ± 3	2	+	4°-20°	Pl. Pr. Ir.	Twins common	ALBITE
1.534	0.001	1 (2)	±	0°	Pl. Pr.	In cavities	apophyllite
1.535	0.010	2	-	0°	Pl. Pr.	Sillimanite incl.	CORDERITE*
1.535	0.019	2	+	0°-45°	Rad. Pl.	Dispersion. H = 3	hydargillite
1.537	0.006 ± 5	2	+	Ir.	Cryptocrystalline	CHALCEDONY
1.539	0.005	1	+	0°	Pr. Ir.	Alters easily	NEPHELITE
1.540	0.009	2	-	0°-15°	Pl. Pr.	Twinning	OLIGOCASE
1.544	0.006	2	±	0°	Pl. Ir.	In pegmatite	eudymite
1.544	0.009	1	+	Wavy 0°	Ir.	Clear	QUARTZ
1.549	0.010 ± 8	1	+	0°	Pr. Pr.	Incl. common	Wernerite
1.549	0.007	2	+	0°-27°	Pl. Pr. Ir.	Twinning	ANDESINE
1.555	0.008	2	+	0°-37°	Pl. Pr. Ir.	Twinning	LABRADORITE
1.56 ±	0.004	2	-	58°	Pl.	In pegmatite	epididymite
1.561	0.008	2	-	0°-45°	Pl. Pr. Ir.	Twinning	BYTOWNITE
1.568	0.008	2	-	0°-20°	Pl. Ir.	Dusty. H = 2	KAOLINITE
1.572	0.020	1	+	0°	Tri. Ir. etc.	H = 4	Alunite
1.575	0.012	2	+	0°-45°	Pl. Pr. Ir.	Twinning	Anorthite
1.57 ±	0.001	2	±	0°	Pl. Ir.	Ultra-blue interf. H = 2	CELORITE*

* Sometimes colored.

TABLE I B.—ANISOTROPIC COLORLESS MINERALS WITH LOW INTERFERENCE COLORS.—(Continued)

Alpha	Biref.	Optic axes	Optic sign	Ext. angle	Shape	Remarks	Mineral
1.578	0.005	1	—	0°	Pr. Ir.	In pegmatites	Beryl*
1.579	0.018	2	—	17°	Pr. Ir. Pr.	In pegmatite, twins	amblygonite
1.61	0.003	1	+	0°	Ir.	In alkaic rocks	eudialite
1.62	0.006	1	—	0°	Ir.	In alkaic rocks	eucolite
1.619	0.010	2	+	0°	Pl. Ir. Pr.	In granites, etc.	Topaz
1.620	0.20	1	—	0°	Ir.	In pegmatite, etc.	TOURMALINE*
1.621	0.014	2	—	0°-32°	Pr. Ac. Pl.	In meta. rocks	wollastonite
1.624	0.009	2	+	0°	Rect.	In limestone	celestine
1.626	0.005	1	—	0°	Pl. Pr.	In ultra-basic rocks	Mellite*
1.63 ±	0.013	2	—	0°	Pl.	H = 4. Brittle	margarite
1.632	0.011	2	—	0°	Pl. Ir.	Graphite incl.	ANDALUSITE*
1.634	0.004 ±	1	—	0°	Ir. Pr.	As inclusions, etc.	APATITE
1.635	0.006	2	—	0°	Ir.	In pegmatite	danburite
1.636	0.011	2	+	0° or 45°	Rhom. Pr.	H = 3	Barite
1.638	0.011	1	—	0°	Ir.	In contact limestone	gchlenite
1.642	0.020 ± 5	2	+	0°	Pr. Ac.	In schist, etc.	SILLIMANITE
1.654	0.016	1	+	Ir.	In pegmatite	phenacite
1.66 ±	0.016 +	2	+	0°	Ir.	In pegmatite	spodumene
1.665	0.009	2	+	0°	Pr.	In basic rocks	ENSTATITE
1.667	0.011	2	+	Ir.	Pseudo-isom.	boracite
1.682	0.009	2	—	Inclined	Ir. Pr. Pl.	In cavities	axinite*
1.698	0.006	2	+	0°	Pr.	Ultra-blue interf.	ZOISITE
1.712	0.016	2	—	0°-45°	Pr. Ac.	In schists	KYANITE*
1.717	0.002	1	±	0°	Ir. Pr.	In contact rocks	vesuvianite*
1.745	0.009	2	+	Ir.	chrysoberyl
1.760	0.008	1 (2)	—	0°	Ir. or Rh.	Parting	Corundum
1.77	0.016	2	+	0°-12°	Ir. Pl.	In schists; twins	Ottrelite*
1.80 ±	0.001	1 (2)	+	0° or 45°	Ir.	Mostly isotropic	GARNET*
1.82 ±	0.004	2	+	Ir.	cerite
1.926	0.016	1	+	Ir.	In pegmatite and veins	scheelite

* Sometimes colored.

TABLE I C.—ANISOTROPIC COLORLESS MINERALS WITH MEDIUM INTERFERENCE COLORS.—(Continued)

Indices	Shape	Sign	Remarks	Mineral
1.55 — 1.73	Pl. Ir. Pr. Ac. Pr.	— + — —	In bentonite and gouge From desert crusts In pegmatite In contact rocks Metam. and peg. Uni.	leverrierite coemanite bertrandite TREMOLITE TOURMALINE*
1.55 — 1.73	Ac. Pr. Ir. Rect. Rh.	— — + + +	In contact rocks Carbonaceous incl. Twinning. In marble Radial. In basic rocks H = 3. In veins and limestone	wollastonite ANDALUSITE Chondrodite pectolite Barite
1.55 — 1.73	Pr. Ir. Ir. Pl. Ac. Pr.	+ + + + +	Rad. In cavities. Dispersion Limestone contacts In schist. 0° ext. Met. rocks. 0° ext. In schist and gneiss	Prehnite humite Anthophyllite SILLIMANITE kornelupine
1.55 — 1.73	Ir. Pr. Pr. Pr. Ir.	— + ± + +	No distinct cleavage In pegmatite. 25° ext. Alters to serpentine In schists Uniaxial, in zinc ore	datolite spodumene OLIVINE lawsonite willemite
1.55 — 1.73	Pr. Ac. Pr. Pl.	+ — +	Ext. 0–38° Ext. 30°, in schist Clinzoisite variety Quartz and graphite incl.	DIOPSIDE KYANITE EPIDOTE Ottrelite
1.55 — 1.73	Pr. Ir. Pl. Ir.	— + + + +	Ext. 32° and 44° Quartz inclusions Indices 1.92 to 1.97. Uni.	rhodonite STAUROLITE* diaspore ZIRCON

* Usually colored.

TABLE I D.—ANISOTROPIC COLORLESS MINERALS WITH HIGH INTERFERENCE COLORS
Birefringence 0.030 and Up
In Oils

Indices			Shape	Axes	Sign	Remarks	Mineral
16	17	18	19	2	—	Cooling saline taste, $H = 2$	Niter
16	17	18	19	1	—	Cooling saline taste, $H = 2$	soda niter
16	17	18	19	2	—	Small optic angle, $H = 1$	TALC
16	17	18	19	2	—	Igneous and metamorphic rocks $H = 2$	MUSCOVITE
16	17	18	19	2	—	Igneous and metamorphic rocks $H = 2$	Lepidolite
16	17	18	19	2	—	$H = 2$. Radial.	pyrophyllite
16	17	18	19	1	—	Symmetrical ext. $H = 3$	CALCITE
16	17	18	19	1	—	Symmetrical ext. $H = 4$	MAGNESITE
16	17	18	19	1	—	Symmetrical ext. $H = 4$	DOLOMITE
16	17	18	19	1	—	Symmetrical ext. $H = 3$	Siderite
16	17	18	19	2	—	0° extinction $H = 4$. Negative elongation	Aragonite
16	17	18	19	2	+	0° extinction $H = 3$	ANHYDRITE
16	17	18	19	2	+	0° extinction. Positive elongation	pectolite
16	17	18	19	2	+	Radial, cavity filling	Prehnite
16	17	18	19	2	—	In basic rocks	datolite
16	17	18	19	1	—	Metamorphic rocks and pegmatites	TOURMALINE*
16	17	18	19	2	+	Igneous rocks, alters to serpentine	OLIVINE
16	17	18	19	2	+	$H = 4$	diaspore
16	17	18	19	2	+		scorodite*
16	17	18	19	2	+	$2V = 16^\circ \pm$	Monazite*
16	17	18	19	1	+		ZIRCON
16	17	18	19	2	+	$2V = 37^\circ$, dispersion	TITANITE*
16	17	18	19	2	—	Multiple twins. Indices over 2.0	baddeleyite

* Usually colored.

TABLE I E.—ANISOTROPIC COLORED MINERALS WITH LOW INTERFERENCE COLORS
Birefringence Up to 0.02, Biaxial unless Otherwise Noted in Remarks
In Oils

Birefringence	Index Beta	Shape	Pleochroism	Color	Remarks	Mineral
0.005 ± 5	1.56	Ir. Ac.	Slight	G. y.	0° ext.	SERPENTINE
0.005 ± 5	1.51	Ir.	None	BR. or blk.	Cryptocrystalline	hisingerite
0.009	1.535	Ir.	Variable	b. g.	Halos around zircon	Cordierite*
0.005 ± 5	1.56	Ir.	Variable	G. b. y.	Ultra-blue interf.	CHLORITE
0.006	1.609	Pl. Ir.	Slight	pink. y.	eudialite*
0.005	1.629	Ir.	Rare	y. br.	Uniaxial	Melilite*
0.002 + ?	1.58	Pl.	Variable	Br. g.	See next table	BIOTITE
0.003 ± 2	1.633	Ir.	Rare	g. b. br.	Uniaxial	APATITE*
0.012	1.649	Ir.	Slight	r. br.	In pegmatite	mosandrite
0.009 ± 4	1.650	Pr.	None	g. colorless	ENSTATITE*
0.009	1.672	Ir.	Fair	Br. v.	In contact rock and veins	axinite*
0.003 ± 3	1.66 ± 6	Ir.	Variable	br.	Variable optically	Allanite
0.01 ±	1.69 +	Ir.	None	bl. r. br. y.	Uniaxial	thorite
0.005	1.697	Pr.	Strong	Pink	thulite
0.007	1.698	Pr.	Slight	y.	In nephelite rocks	hiortdahlite
0.005	1.703	Pr.	Fair	b. g.	Multiple twins	serendibite
0.003 ± 2	1.703	Ir.	Slight	br. g.	Uniaxial, zoned	Vesuvianite*
0.006	1.709	Ir.	Weak	b. g.	In schist and gneiss	sapphirine
0.012 ± 4	1.75	Pl.	Strong	gray, g.	In schists. Has conclusions.	Chloritoid
0.008	1.759	Ir.	Slight	r. b. etc.	Uniaxial	Corundum.
0.002 ± 1	1.70	Ir.	None	R. br. g.	Zoned and twinned	GARNET
0.006	1.80	Pr.	Strong	br.	enigmatite
0.004	1.817	Ir.	Slight	br. r. gray	In gneiss	cerite*

* Usually colorless.

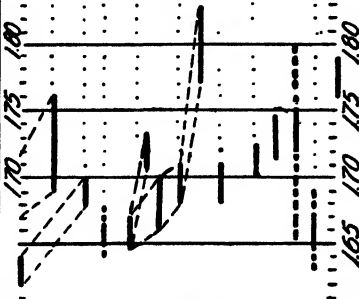
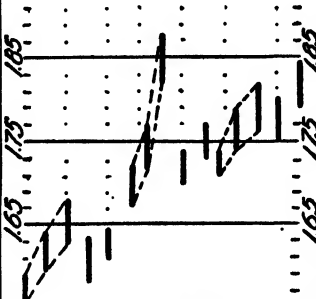
TABLE I F.—ANISOTROPIC COLORED MINERALS, MEDIUM INTERFERENCE COLORS
Birefringence 0.01 to 0.03. Biaxial Except as Noted

In Oils

Interf. color	Indices and birefringence	Shape	Pleo- chromism	Color	Remarks	Mineral
Low to medium		Ir. Ac.	Slt.	G. y.	0° ext. Uniaxial	SERPENTINE
		Pl. Ir.	Slt.	y. y.	(-)	meliphanite
		Ir. Pl.	Fair	G. y.	0° ext.	CHLORITE
		Ir. Pr.	Var.	rose, br.	In pegmatite, etc.	ANDALUSITE*
		Ir.	Str.	b. y. r.		dumortierite
		Pr. Ir.	Slt.	y.	In nephelite rocks	hainite
		Pr.	Slt.	g. pink, br.	0° ext. (+)	BRONZITE
		Pr.	Fair	G. R. br.	0° ext. (-)	HYPERSTHENE
		Pl.	Slt.	b. g. y. pink	In pegmatite	triphylite
		Ir.	Fair	b. g. y. pink	In pegmatite	lithiophyllite
		Pr. Ac.	Str.	B.	0°-30° ext.	KYANITE
		Ir.	Slt.	y.	Uni. (-)	melanocerite
		Pl.	Fair	gr. g.	In slates, graphitic	Ottrelite
		Pl.	Str.	gr. g.	In altered rocks	Chloritoid
Medium		Pr.	None	Y.	Uni. with nephelite	Cancrinite*
		Ir.	Fair	y. y.	As a clay	nontronite
		Ir.	Str.	y. g.	In pegmatite	hercynite
		Ir.	Fair	Azure	X colorless	lazulite
		Pr. Ir.	Str.	b.	Y colorless	grandierite
		Ir. Pr.	Slt.	b. br. y.	Twinning	catapleite
		Ir. Pl.	{ None Fair	G. br.	Fibrous or in granules	Glaucophane
		Ac.	Var.	pale g.	0°-25° ext.	Actinolite
		Pr. Ac.	Str.	G. br. b.	2 cleavages, 56°, 0°-25° ext.	HORNBLÉNDE
		Pr.				

* Usually colorless.

TABLE I F.—ANISOTROPIC COLORED MINERALS, MEDIUM TO HIGH INTERFERENCE COLORS.—(Continued)

Interf. color	Indices and birefringence	Shape	Pleo- chromism	Color	Remarks	Mineral
Medium		Pr.	Fair	LAVENDER, b.	About 5° ext.	Glaucophane
		Pr.	Str.	BR.	0°-8° ext.	Basaltic
		Pr. Ac.	Str.	B. blk.	12°-15° ext.	hornblende
		Pr.	None	y. br. g.	0° ext.	Arfvedsonite
		Pr.	Var.	g. pink, br. gr.	2 cl. at 89°, 0-42° ext.	SILLIMANITE*
		Pr.	Slt.	g. br.	3 cleavages, 0-42° ext.	AUGITE
		Pr.	Var.	G. b. y.	2 cl. at 89°, 0-30° ext.	DIALLAGE
		Pr.	Fair	G. b. y.	2 cl. at 89°, 0-8° ext.	Acmite-augite
		Pl. Ir.	Slt.	y.	In alkalic rocks	Acmite
		Ir.	Slt.	y. br.	In syenite	rosenbuschite
		Pr.	Var.	BR. dark g.	Dispersion	woehlerite
		Ir.	Str.	BR. B. gr. br. pink	Uniaxial	babingtonite
Medium high		Pl.	Str.	light BR. }	Basal plates show no pleoch.	Phlogopite
		Pl.	Str.	BR. g. }	and low interf. Small optic	BIOITTE
		Pl.	Str.	dark BR.	angle	Lepidomelane
		Pr. Pl.	Str.	b. g. v.	Water deposits	vivianite
		Ir.	Fair	br. y.	Chondrodite*
		Ir.	None	g. y.	OLIVINE*
		Ir. Pr.	Fair	BR. g. y.	Fayalite
		Pl.	Var.	BR. r.	From iron silicates	Iddingsite
		Pl. Pr.	Str.	br. y.	In alkalic rocks	Astrophyllite
		Ir. Pr.	Var.	Y. G. br.	0° to 36° ext.	EPIDOTE
		Ir.	Str.	y. r. v.	pyrochroite
		Ir.	Fair	b. v.	Uniaxial (+)	benitoite
Pl.	None	Y. r. br.	4° ± ext. 2V = 16°	Monazite		

* Usually colorless.

TABLE I F.—ANISOTROPIC COLORED MINERALS, HIGH INTERFERENCE COLORS.—(Continued)

Interf. color	Indices and birefringence	Shape	Pleo- chromism	Color	Remarks	Mineral
Very high	<div>Uniaxial</div> <div>Biaxial</div>	Rh.	Sit.	gr. br. y. }	The carbonate group H = 3 to 4. Causes halos Zones and twins Nearly octahedral in small grains Translucent H = 2 0° ext. Dispersion, 2V = 36° H = 2 12° ext. 17° ext. 0° ext. Translucent. In pegmatite Translucent Dispersion	{ CALCITE* DOLOMITE* SIDERITE* ZIRCON* Cassiterite Rutile octahedrite HEMATITE copiapite xenotime ilvaite TITANITE Sulphur baddeleyite* wolframite Goethite columbite Manganite Brookite
		Rh.	Sit.	gr. br. }		
		Ir.	Sit.	pale br.		
		Ir.	Var.	BR. R. gr.		
		Ir. Ac.	Sit.	BR. y.		
		Ir. Pl. Pyr.	Sit.	y. b. gr.		
		Ir. Pl.	Sit.	R. br.		
		Ir. Pl.	Sit.	y. br.		
		Pr.	Sit.	v. br.		
		Ir. Pl.	Fair	BR. blk. g.		
		Ir.	Var.	R. BR. y.		
		Ir.	Fair	Y.		
		Pl. Ir.	Fair	br. blk.		
		Pl. Pr. Ir.	Fair	BR. R. g. blk.		
		Pr.	Fair	br.		
		Pl.	Sit.	br.		
		Pr. Pl.	Fair	BR. Y. r. blk.		
		Ir.	Sit.	br. y. r.		

* Usually colorless.

TABLE II A.—OPAQUE MINERALS IN THIN SECTION

Color in reflected light	Shape	Occurrence and remarks	Mineral
Bluish black	{ Grains; octahedrons Lathlike; flakes	Twinning	MAGNETITE
Dark lead gray	Massive	With leucoxene; submetallic	ILMENITE
Black to steel gray	{ Grains; dust; amorphous Crystals	In ores; gray streak	chalcocite
Lead gray	Massive	Metamorphic rocks; rarely metallic	GRAPHITE
Black to brown black	Grains; octahedrons	Sediments and weathered rocks; metallic to dull	pyrolusite
		Shows triangular pits; alters to cerussite or anglesite	Galenite
Bronze yellow	Irregular; tabular crystals	In ferromagnesian rocks; brownish red to black in transmitted light	
Brass yellow	Isometric forms	In magmatic segregations; brown; alters to pyrite	CHROMITE
Greenish yellow	Irregular; tetragonal	PYRRHOTITE
Pale yellow	Grains; irregular	In ores; green black streak	PYRITE
Yellow or brown	Massive and earthy	Pale bronze yellow; often alters to pyrite	CHALCOPYRITE
		Stainings in sediments about iron minerals; pseudomorphs after ferruginous minerals	Marcasite
Copper red to brown	Massive	In ores; tarnish iridescent	LIMONITE
Red	{ Scales; rhombohedrons Irregular crystals	In weathered rock originally containing iron; dusty indian red to black in transmitted light	bornite
Silver white	Subhedral to euhedral gr.	Metallic copper red	HEMATITE
	{ Sealy; dusty; aggregates Fine grained	In ores; in crystalline rocks	Copper
White to gray	Fine grained	Sediments and in altered feldspars, often brown in transmitted light	arsenopyrite
	{ Fine grained Octahedral	In hydrothermally altered rocks; bright luster; on iron minerals	KAOLIN
		In hydrothermally altered rocks; bright luster; on ilmenite	SIDERITE
		In slates	LEUCOXENE
			Octahedrite

TRANSLUCENT MINERALS (See other tables)

Goethite.	Rutile.
Ilvaite.	Sphalerite.
Manganite.	Turgite.
	Wolframite.

TABLE II B.—ISOTROPIC MINERALS IN THIN SECTION
Index Less than Balsam

Relief	Color	Cleavage	Remarks	Mineral
High.....	Colorless Colorless, blue, yellow, etc. Always colorless	None 4 good None	As inclusions, color fringes, shape rounded In veins and igneous rocks Cavity fillings, always secondary Matrix for crystals	GAS and LIQUIDS Fluorite OPAL
Medium.....	Colorless, yellow, brown Colorless, blue, yellow, pink Blue, brown, etc., irregularly distributed Blue, irregularly distributed Orange to dark brown Blue	None 6 poor 4 poor 4 poor None	Usually in large crystals in alkalic rocks low in silica; inclusions regularly arranged in sodalite and in irregular zones in haunynite and noselite	SILICIC GLASS SODALITE
Low.....	Colorless Colorless, gray White to yellow White to yellow Colorless Varied	Poor None None None 1 perf. None	Finely crystalline, amorphous In contact limestones No regular inclusions; cavity filling or groundmass Rounded isometric forms; inclusions in zones Amorphous, earthy, opaline Groundmass for crystals Cryptocrystalline; biref. very weak	Haunynite Noselite hisingerite Lazurite ANALCITE LEUCITE halloysite Andesite glass apophyllite* Chalcedony
Index Greater than Balsam				
Low.....	Colorless Orange to dark brown Usually brown	Good None None	Cubic form and cleavage Alteration and vein filling Matrix for crystals	periclase hisingerite Basic glass
Medium.....	Brown or green Green or blue Green to black	Not seen Not seen None	Confused aggregates Granular in iron rocks	Biotite* Chlorite* Greenalite
High.....	Green, red, brown, blue Colorless, pink, brown, etc. Brown to black, red Brown, yellow, etc.	None None None Poor	In basic rocks Isometric shapes Isometric, nearly opaque Igneous accessory	Spinel GARNET CHROMITE Allanite
Very high.....	Brown, green, yellow, color- less Red, brown to colorless Colorless, etc.	6 good None Octahedral perfect	Larger crystals anisotropic In peridotite	sphalerite Perovskite Diamond

* Generally anisotropic

TABLE II C.—CONFUSED AGGREGATES IN THIN SECTION

Color	Remarks	Mineral
White to gray.....	Radiating or parallel fibers $n = 1.537$	Chalcedony
White to gray.....	Platy to fibrous. $\alpha = 1.552$. High } interf. In potash rocks, $2V = 45^\circ \pm$ }	SERICITE
White to gray.....	Soda mica occurs in sodic rocks	Paragonite
White to gray.....	Occurs in magnesian rock. Platy to } fibrous. $\alpha = 1.537$, $2V = 10^\circ \pm$. } High interf.	TALC
White to gray.....	In minute rhombs or fibrous and radiating	CALCITE
Gray opaque.....	Dull, dust; $n = 1.563$. Birefringence = 0.006	KAOLINITE
Gray opaque.....	Brilliant in reflected light	EPIDOTE
Gray opaque.....	Brilliant in reflected light. Occurs in basic igneous rocks, often on ilmenite	Leucoxene
Gray opaque.....	Brilliant in reflected light, hydrothermal	Siderite
Yellowish brown to nearly opaque.....		LIMONITE
Brownish dust.....	White in reflected light, $n = 1.563$, Biref. = 0.006	KAOLINITE
Brownish fibers.....	White in reflected light, $n = 1.537$	Chalcedony
Yellowish green to green.....		EPIDOTE
Blue green.....	Medium interference colors. Altered olivine	SERPENTINE
Blue green.....	Medium interference colors. Altered hornblende	Uralite
Blue green.....	Almost isotropic	CHLORITE
Mixed.....	Altered calcic feldspar in basic rocks. Principally zoisite	Saussurite
Black.....	Opaque black dusty areas in igneous rocks; magnetite, etc.	Opacite

Notes on Troublesome Minerals in Thin Sections.—1. *Nephelite* resembling quartz and orthoclase. The mineral is often first suspected from the field occurrence or from a microscopic association with acmite, riebeckite or a feldspathoid in igneous rocks. It does not occur with quartz. It has better cleavage than quartz but poorer than orthoclase. The bars of the interference figure are so wide as to make its nature uncertain. The index is very close to that of balsam, whereas quartz is higher and orthoclase lower than balsam. The forms of euhedral grains (and pattern of inclusions in large anhedral grains) are both rectangular and hexagonal, varying with the position of the section. The alteration is likely to be different from that of other minerals. Nephelite is (—), quartz is (+). Nephelite gives a silica jelly, and in the absence of other minerals that give a jelly the test is very useful.

2. *Sericite, Talc, Paragonite, Leverrierite*.—The occurrence is most useful; sericite in rocks with feldspar, especially potash feldspar; paragonite in sodic rocks; talc in magnesian rocks; leverrierite in rocks deficient in alkali, like altered clays. Leverrierite has a birefringence from 0 to 0.030, which is lower than the others. The lowest index, α , is 1.539 for talc and 1.561 for sericite, but the range of indices for the two minerals shows overlap.

3. *Olivine from Diopside*.—Diopside has two cleavages, and olivine rarely one poor one. Diopside has extinction angles up to 45° , olivine 0° . Olivine is almost invariably altered and alters more readily than diopside, forming green serpentine along a network of curving fractures.

(Continued on page 469)

TABLE II D.—ANISOTROPIC COLORLESS MINERALS WITH LOW INTERFERENCE COLORS IN THIN SECTION

Beta	Cleavage	Ext. angle	Optic axes	Optic sign	Elong.	Remarks	Mineral
1.48	1 poor	2	-	...	No regular inclusions	Analcite
1.48	1 poor	0	2	+	-	Gas inclusions, wedged twins	Tridymite
1.48	1 good	Var.	1-2	±	±	Radiating fibrous, amygdules	Zeolites
1.48	Var.	1 (2)	-	Cavities in volcanic rock	Cristobalite
1.48	1 good	0	2	+	+	Amygdaloidal	natrolite
1.49	1 good	0-5	2	-	-	Sheaflike aggregates	stilbite
1.50	1 good	0-6	2	+	-	Cavities, basalts and schists	heulandite
1.50	None	2	+	Regular incl., rounded grains	LEUCITE
1.52	2 perf.	0-5	2	-	+	Carlsbad twins or none, dusty	ORTHOCLASE
1.52	2 perf.	0-15	2	-	-	Gridiron twinning, dusty	MICROCLINE
1.52	1 g., 2 p.	0-52	2	+	±	Arrowhead twins	Gypsum
1.53	2 perf.	0-20	2	+	+	Albite twinning	ALBITE
1.53	1 perf.	0	1 (2)	±	Cavity filling	apophyllite
1.53	1 poor	0	1	-	Sodic rocks	NEPHELITE
1.54	1 poor	0-30	2	-	-	Sillimanite incl.: often twinned	Cordierite*
1.54	1 good	0-44	2	+	-	Dispersion strong	hydrargillite
1.54	None	0-29	2	?	±	Cryptocrystalline aggregate	CHALCEDONY
1.54	2 perf.	0-15	2	±	±	Albite twins	OLIGOCASE
1.54	1 poor	0	1	+	Inclusions common	marialite
1.55	None	0	1 (2)	+	Arranged inclusions? Clear.	QUARTZ
1.56	1 perf.	0-20	2	-	+	Dusty to nearly opaque	KAOLINITE
1.57	Prismatic	0	2	+	+	Fibrous	SERPENTINE*
Var.	2 perf.	0-45	2	±	±	See Table III.	PLAGIOCLASE
1.57	?	0	1	±	beryl*
1.58	1 good	0	1	+	-	Tabular or cubic	Alunite
1.59	1 perf.	0-3	2	+	-	Fine twinning bands possible	Chinolore*
Var.	Fibrous	Var.	2	±	±	May be platy.	CHLORITE*
1.62	1 perf.	0	2	+	±	Incl., scarcity of twinning	Topaz
1.62	1 perf.	0	2	+	+	In limestone	celestite
1.63	2 perf.	0-32	2	-	±	In contact rocks. Needles	wollastonite

* Sometimes colored. * Usually colored.

TABLE II D.—ANISOTROPIC COLORLESS MINERALS WITH LOW INTERFERENCE COLORS IN THIN SECTION.—(Continued)

Beta	Cleavage	Ext. angle	Optic axes	Optic sign	Elong.	Remarks	Mineral
1.63	None	0	1 (2)	-	(+)	Hexagonal prisms	APATITE*
1.63	1 poor	1	±	(±)	Incl. in it are pegs.	melilite*
Var.	None	0	1 (2)	-	-	Triangular needles	TOURMALINE*
1.64	2 good	2	-	-	Graphite incl.	ANDALUSITE
1.64	1 good	0-6	2	-	+	Metamorphic rocks	margarite
1.64	2 perf.	0-45	2	+	+	Twinning common	Barite
1.65	2 good	0	2	+	+	Incl. elongated parallel c axis	ENSTATITE
1.66	1 good	0	2	+	+	Needles, nearly square prisms	SILLIMANITE
1.67	None	0	2	+	(+)	Complex twinning	boracite
1.67	7 poor	0	1	-	..	Contact mineral	gehlenite
1.68	3 good	Var.	2	-	..	Wedges, alters to chlorite	axinite*
1.69	1 good	0	2	+	±	Prismatic, ultra-blue interf.	ZOISITE
1.72	1 + good	0-30	2	-	+	Long-bladed crystals, in schist	KYANITE
1.72	Poor	0	1 (2)	-	+	Zonal arrangement of colors	Vesuvianite
1.74	1 poor	0	2	+	+	Often twinned	STAUROLITE*
1.74	Poor	2	+	chrysoberyl
1.75	None	1 (2)	-	..	May be partly isotropic, zoned	GARNET*
1.76	? none	1 (2)	-	..	Parting at 94°	Corundum
1.77	1 perf.	0-25	2	+	..	Organic matter arranged	Ottrelite*
1.82	None	2	+	..	In gneisses	cerite

* Sometimes colored.

* Usually colored.

TABLE II E.—COLORLESS ANISOTROPIC MINERALS WITH MEDIUM INTERFERENCE COLORS IN THIN SECTION

Beta	Biref.	Sign	Cleavage	Extinction	Elongation	Remarks and occurrence	Mineral
1.480	0.012	+	Needles	0°	+	Secondary in cavities	natrolite
1.507	0.039	+	Needles	+	Fib. in amygdaloid; uniaxial	thauwasite
1.508	0.028	-	Needles	0°	±	Secondary in cavities	THOMSONITE
1.524	0.028	-	1 good	0°	±	Uniaxial; in alkalic rocks	Cancrinite
1.524	0.012	-	Needles	20°-40°	+	In cavities with basic rocks	Laumontite
1.534	0.027	+	Needles	0°	+	Radiating; veins; slates	wavellite
1.540	0.013	+	Needles	0°	+	Fibers in metamorphic rocks	Chrysotile
1.561	0.021	+	1 perf.	21°-44°	+	Uniaxial in altered Mg rocks	brucite
1.564	0.030	+	1 perf.	0°	+	In metam. magnesium rocks	PHLOGOPITE*
1.566	0.023	+	1 good	25°	-	Alteration from Al rocks	hydrargillite
1.567	0.022	-	1 fair	In contact rocks; uniaxial	Wernerite
1.572	0.020	+	1 fair	0°	-	Uniaxial; from hydrothermal alteration of feldspars	Alunite
1.576	0.044	+	3 good	0°	..	Cleavages at right angles	Anhydrite
1.584	0.012	-	2 good	37°	..	Multiple twinning	Anorthite
1.589	0.050	-	1 good	0°	+	In metam. and Mg rocks	TALC
1.59 ±	0.011	+	1 perf.	2°-7°	-	In schists and altered rocks	CHLORITE*
1.590	0.038	-	1 perf.	0°	+	In silicic igneous rocks, schists	MUSCOVITE
1.602	0.044	-	1 perf.	0°	+	Schists, gouge, sediment, clays	leverrierite
1.606	0.038	+	1 perf.	0°	+	In cavities in basic rocks, radial	pectolite
1.623	0.027	-	Fibrous	25°-18°	+	Tufted needles, Cl at 56°	TREMOLITE
1.624	0.035	+	2 good	0°	..	Calcium contact rocks	humite
1.626	0.033	+	1 poor	0°	±	In cavities, radial. Dispersion	Prehnite
1.62 ±	0.012	±	1 perf.	0°	±	In schist and propylite	CHLORITE*
1.629	0.015	-	1 + perf.	0°-32°	±	In metamorphosed limestone	wollastonite
1.637	0.012	+	1 + perf.	0°-45°	±	In sediments and veins	Barite
1.638	0.011	-	1 good	0°	-	Carbonaceous inclusions, in metamorphic Al rocks	ANDALUSITE
1.63 ±	0.031	+	None	28° ± 3	..	Contact-metamorphic rocks	Chondrodite
1.642	0.024	+	2 perf.	0°	+	In schists, needles	Anthophyllite

* Usually colored.

TABLE II E.—COLORLESS ANISOTROPIC MINERALS WITH MEDIUM INTERFERENCE COLORS IN THIN SECTION.—(Continued)

Beta	Biref.	Sign	Cleavage	Extinction	Elongation	Remarks and occurrence	Mineral
1.653	0.044	-	None	0°-4°	..	In cavities with zeolites	datolite SILLIMANITE spodumene DIOPSIDE lawsonite Diallage kornepine OLIVINE TOURMALINE*
1.659	0.022	+	1 good	0°	+	In metamorphic rocks	
1.666	0.016	+	2 good	25°	..	Mostly in coarse pegmatites	
1.671	0.030	+	2 perf.	0°-38°	±	Occurrence variable	
1.674	0.019	+	2 good	0°	..	In amphibole schists	
1.675	0.029	+	2 + part.	38°	+	In basic crystalline rocks	
1.677	0.012	-	2 good	0°	-	In schists and gneiss	
1.69 ± 4	0.037	±	1 poor	0°	..	Not with quartz. Alters to serpentine mesh	
1.687	0.046	-	None	0°	-	Dark when normal to plane of vibration; triangular prisms; uniaxial	
1.694	0.029	+	2 + poor	Contact zinc ores; uniaxial	
1.720	0.016	-	2 good	0°-32°	+	In metamorphosed Al rocks	KYANITE diaspore rhodonite ZOISITE STAUROLITE* ZIRCON*
1.722	0.048	+	1 + perf.	0°	-	In schists, volcanic rocks	
1.735	0.011	±	2 perf.	32°-44°	..	In Ca and Mn contact rocks	
1.736	0.012	+	1 perf.	0°-21°	..	Ultra-blue interf. Result of metam.	
1.750	0.012	+	1 poor	0°-45°	..	Some X-twins; only in crystalline metamorphic rocks	
1.94 ± 2	0.051	+	? none	0°	..	Uniaxial; in igneous and metamorphic rocks	

* Usually colored.

TABLE II F.—COLORLESS ANISOTROPIC MINERALS IN THIN SECTION
Part 1. Moderately High Interference Colors

Axes	Sign	Biref.	Elong.	Relief	Cleavage	Appearance	Remarks	Mineral
1	—	0.030	—	Low	1 fair	Grains, lathlike, prisms	Alters to kaolin or muscovite.	Wernerite
2	—	0.048	±	Low	1 perf.	Massive, fibrous	metamorphic rocks	pyrophyllite
2	—	0.050	±	Low	1 perf.	Scaly	A clay, radial	TALC
2	±	0.06 ±	..	Low	? perf.	Lamellar, massive	In magnesian rocks	copiapite*
2	±	0.06 ±	..	Med.	1 perf.	Fibrous, divergent	Variable	vivianite*
2	—	0.086	+	Med.	1 perf.	Scales, shreds	Parallel extinction	MUSCOVITE
1	—	0.021	—	Med.	None	Prismatic, radiating	Parallel extinction	TOURMALINE*
2	+	0.033	+	Med.	1 poor	Tabular, fan shaped	Inclusions of chlorite, in veins and amygdaloids	Prehnite
2	+	0.043	—	Med.	3 perf.	Prismatic	Alters to gypsum, parallel extinction	Anhydrite
2	—	0.044	+	Med.	None	Prisms	With zeolites in basic igneous rocks	datolite
2	±	0.032	+	High	Poor	Variable	scorodite*
2	±	0.036	±	High	1 poor	Prismatic	Alters to serpentine	OLIVINE
2	±	0.038	±	High	1 perf. fibrous	Acicular	Mostly in veins, 0° ext.	pectolite
2	+	0.048	—	High	1 perf.	Prismatic, flattened	Assoc. with corundum	diaspore
2	+	0.050	±	High	1 + good	Prismatic, angular	2V = 15° ±	Monazite*
1	+	0.062	..	Very high	? none	Short prisms	Parallel extinction.	ZIRCON
2	—	0.07	..	Very high	1 perf.	Tabular	Often polysynthetic twins	baddeleyite

Part 2. Very High Interference Colors

Axes	Sign	Biref.	Elong.	Relief	Cleavage	Appearance	Remarks	Mineral
1	—	0.172	..	Var.	Rh.	Grains, fibrous, oölitic	Multiple twinning common	CALCITE
1	—	0.172	..	Var.	Rh.	Rhombs	DOLOMITE
1	—	0.202	..	High	Rh.	Grains	No polysynthetic twins	Magnesite
1	—	0.240	..	High	Rh.	Alters to limonite	Siderite
2	—	0.155	..	High	Poor	Acicular	Polysynthetic twinning	Aragonite
1	+	0.287	..	High	None	Grains and Acic.	Parallel extinction	Rutile*
2	+	0.121	..	Very high	Poor	Wedge shaped	2V = 37°, dispersion	TITANITE*

* Often colored.

TABLE II G.—ANISOTROPIC COLORED MINERALS IN THIN SECTION
Low Relief. Indices 1.532 to 1.565

Birefringence	Color	Pleochroism	Cleavage	Axes	Extinction	Elongation	Remarks	Mineral
0.006— 0.010 0.009 0.009 0.044 0.064 ± 6	b.g.y.br. G. br.-r. br.-y. BR.g. Y.r.v.	Slight Slight Weak Fair; pale colors X = y - g; Y = pale y Z = sulfur y.	1 poor 1 perf. ? Perf.	2 - 2 - 2 - 2 +	X = c 0° ± X = c	- + + ..	Oriented graphite inclusions Pseudomorphous Fibrous 2V = 0° to 25° May show inclined extinction	Cordierite* SERPENTINE Antigorite Phlogopite copiaite
Medium Relief. Indices 1.581 to 1.700 +								
± 0.005 ± 0.004 0.005 0.006 0.010 0.011 0.012 0.012 ± 0.015 0.018 0.018 0.020	blk. br. G.b. b.g. y.g. y. pink BR.v.b.y. r.br.y.g.b. g. R.G. g.y.o. Pink G.y. G. Distinct X = Y = g. Usually none X = y; Z = r Variable X = y; Y = v - b; Z = colorless Slight X = r; Y = y; Z = g, b X = y - br; Z = y X = r; Y = pk; Z = y X = y; Y = Z = g, y Z = g, b; Y = X = g	None 1 perf. None Poor Poor 3 good 2 good 2 good 2 good 1 perf. 1 perf. 1 ± 2 perf.	2 - 2 ± 1 - 1 - 1 ± 2 - 2 - 2 + 2 - 2 + 2 - 2 +	0° ± Inclined X = c X = a; Z = b Z = c 0° X near c Z ∠ c = 15°	± - - + + ± + + + + +	Usually amorphous Fibrous, ultra-blue interf. Hexag. needles Anomal. interf. Wedge-shaped, in veins and contact rock Oriented inclusions Oriented inclusions Earthy Fibrous granules 2V is moderate	hisingerite CHLORITE APATITE* Melilita* eudialite* axinite* ANDALUSITE* BRONZITE HYPERSTHENE nontronite thulite Glauconite Pargasite

* Usually colorless.

TABLE II G.—ANISOTROPIC COLORED MINERALS IN THIN SECTION.—(Continued)

Birefrin- gence	Color	Pleochroism	Cleavage	Sign		Extinction	Elongation	Remarks	Mineral
				Axes	Sign				
Medium Relief. Indices 1.581 to 1.700 +									
0.020	b.v.y.	X = colorless; Y = v; Z = b	2 perf.	2	—	Z \wedge c = 5°	+	Metamorphic only	Glaucophane
0.02 ± 1	BR.y. dark	X = v; Z = r. — br. var.	None	2	—	±	Variable	Allenite
0.022	G.br.	X = v; Z = r. — br. var. g.b. and br.	2 perf.	2	±	Z \wedge c = 15° to 25°	±	Cleavage at 124°	HORNBLÉNDE
0.025	BR.y.g.r.B.	X < Z	None	1	—	X = c	—	Prismatic crystals, Dark \perp plane of vib.	TOURMALINE
0.027	G.	Slight	2 perf.	2	—	Z \wedge c = 15°	+	Needles	ACTINOLITE*
0.027	g.br.v.r.	Slight	2 perf.	2	+	Z \wedge c = 42°	AUGITE
0.035	g.br.	None	1 poor	2	±	Alters to serpentine	OLIVINE*
0.036	b.	X = colorless; Y = Z = b	Poor	2	—	Often twinned	lazulite
0.045	BR.r.	X < Y < Z	3 poor	2	—	X = a	..	Pseudomorph after olivine.	Iddingsite
0.050	BR.y.g. r.blk.	Strong X = y Y = r.br. Z = r.br.	1 perf.	2	—	0-3°	+	Dispersion 2V is very small	BIOTITE

Notes.—See also High relief for acmite-augite, arfvedsonite, augite

* Usually colorless

• May be colorless

TABLE II G.—ANISOTROPIC COLORED MINERALS IN THIN SECTION.—(Continued)
High Relief. Indices 1.70 to 2.00

Birefrin- g.	Color	Pleochroism	Cleavage	Axes Sign	Extinction	Elongation	Remarks	Mineral
0.002	R.y.b.g.	None	None	..	Zoned	..	Often isotropic	GARNET*
0.004	br.r.g.	Slight	None	2+	Some biaxial, zonal color	cerite
0.005	Var.	Slight	Poor	1±	Dispersion, multiple twins	Vesuvianite*
0.005	b.y.g.	X = Y = y; Z = b	None	2+	35° to twinning	..	Dispersion	serendibite
0.006	B.g.	Slight	None	2-	..	+	..	sapphirine
0.012±5	g.gr.b.	Strong	1 good	2+	X = b	..	Dispersion, in schists	Chloritoid
0.008	b., etc.	Slight	Partings	1-	..	+	..	Corundum*
0.010	br.r.o.	..	2 poor	1+	Often isotropic	thorite
0.010	BR.y.	Fair	1 poor	2+	Z = c	..	In schists	STAUROLITE*
0.010	g.g.-b.	Fair	1 good	2+	Z ∧ c 25° +	..	In slates	Ottrelite
0.015	b., etc.	Slight	2 good	2-	Z ∧ c 30° +	..	In schists	KYANITE*
0.02±	BR.y.	X = y; Z = r. - br.	Poor	2-	..	±	Variable	Allanite
0.021	B.blk.	Strong	2 perf.	2-	X ∧ c 12°-15°	..	Cl. angle 124°	ARFVEDSONITE
0.025	g.r.	Slight	2 perf.	2+	Z ∧ c 42°	..	Cl. angle 89°	AUGITE
0.029	g.v.	Fair	2 perf.	2+	X ∧ c 6°-33°	-	Cl. angle 89°	ACMITE-AUGITE
0.031	BR.blk.	Fair	2 perf.	2-	Z ∧ c 0°-10°	+	Cl. angle 124°	BASALTIC
0.033	g.br.	X = b.g; Y = br.; Z =	2 perf.	2+	Dispersion	HORNBLende
0.039	Y.G.	br., g. Fair	1 perf.	2-	X ∧ c 5°	..	Metamorphic	babingtonite
0.04±	BR.r.	Var.	3 poor	2+	X = a	..	After olivine	EPIDOTE
0.050±	r.br.y.	Fair	1+ good	2+	Z ∧ c 6°±	..	2V = 16°	Iddingsite
0.055	y.c.r.br.	Strong	1 perf.	2+	Z = b	..	2V = 75°	Monazite
0.05±1	BR.y.g.	None	? none	1+	..	+	..	Astrophyllite
								ZIRCON

* Usually colorless.
• May be colorless.

TABLE II G.—ANISOTROPIC COLORED MINERALS IN THIN SECTION.—(Continued)

Birefrin- gence	Color	Pleochroism	Cleavage	Axes Sign	Extinction	Elongation	Remarks	Mineral
High Relief. Indices 1.70 to 2.00								
0.050	G.b.blk.	Fair	2 perf.	2 —	$X \wedge c 2^\circ-5^\circ$	—	Deep color	Acmite
0.051	BR.y.	Slight	2 poor	2 —	$Z = a$..	Abnorm. blue to yellow interf.	Fayalite (iron olivine)
0.061	PINK r.b.	Strong	1 perf.	2 +	$X \wedge c 7^\circ$	+	Wedges, $2V = 37^\circ$	pidmontite
0.12 ± 1	br.y.gr.	Fair	Poor	2 +	$Z \wedge c 51^\circ$..		TITANITE*
0.232	gr.	Fair	Rhomb.	1 —		Siderite
Str.	br.blk.	Strong	Poor	2 —	$Z = c$..		ilvaite
Very High Relief. Indices 2.00 to 2.71								
0.10 ± 4	R.y.br.	Slight	1 perf.	2 —	0°	..	Nearly opaque	Goethite
0.25 ±	r.blk.	Slight	Parting	1		Hematite
0.061	br.y.b.	Fair	Poor	1 —		Octahedrite
0.070	b.	Fair	1 perf.	2 —	$X \wedge c 12^\circ$	±		baddeleyite
0.096	br.r.y.g.	Fair	2 perf.	1 +	+	Zones + twins	Cassiterite
0.121	r.br.	Slight	Poor	Wedges	TITANITE
0.150	br.blk.	Fair	1 perf.	2 +	Nearly opaque	Wolframite
0.158	br.y.	Fair	None	2 +	$Z = a$	+		Brookite*
0.274	BR.r.	Slight	None	1 +	+	Acicular	Rutile
0.276	y.g.gr.							
0.290	blk.	Fair	3 perf.	2 +	$Y = b$..	Nearly opaque	manganite
0.290	Y.	Var.	Poor	2 +	$X = a$..		Sulphur

* May be colorless.

TABLE III.—FELDSPARS

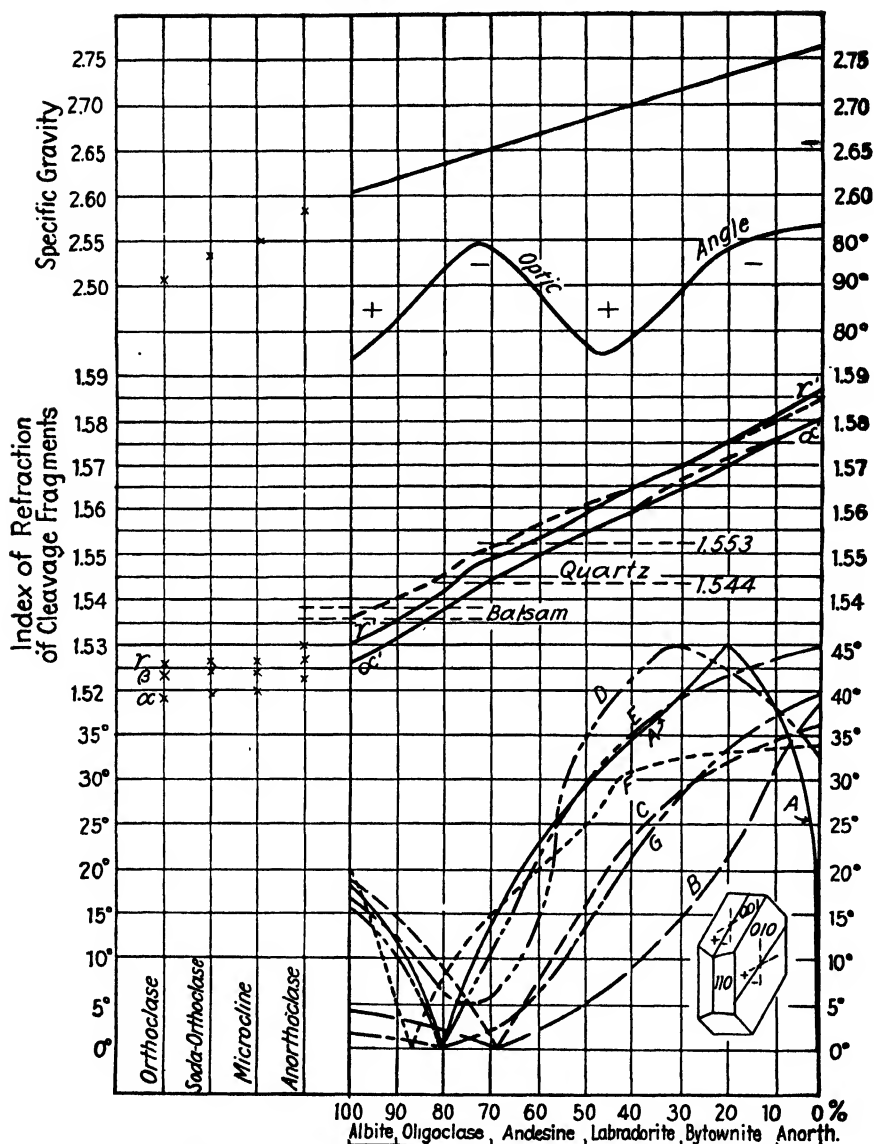


FIG. 264.—Curves for feldspar determinations. Compiled from various sources. Extinction-angle curves without regard to sign. A, Statistical method; maximum symmetrical extinction. B, On cleavage plates parallel to the best cleavage (001). C, On cleavage plates parallel to (010). D, On the rays of spherulites. E, On blocks cut at right angles to both (001) and (010). F, On plates cut at right angles to a bisectrix. G, On plates at right angles to the optic normal (Federow). Note that the angles for oligoclase are low in all positions and that those for bytownite are high in nearly all positions. Note that the feldspars with medium extinction angle occur in two groups with very different indices. Index of refraction curves are based on cleavage fragments. If the lesser index is determined it makes little difference which cleavage rests on the slide. $N <$ balsam for albite, $N >$ balsam for andesine. (See also the curves in *Am. Mineralogist*, vol. 11, pp. 139–154, 1928.)

TABLE III.—FELDSPARS.—(Continued)

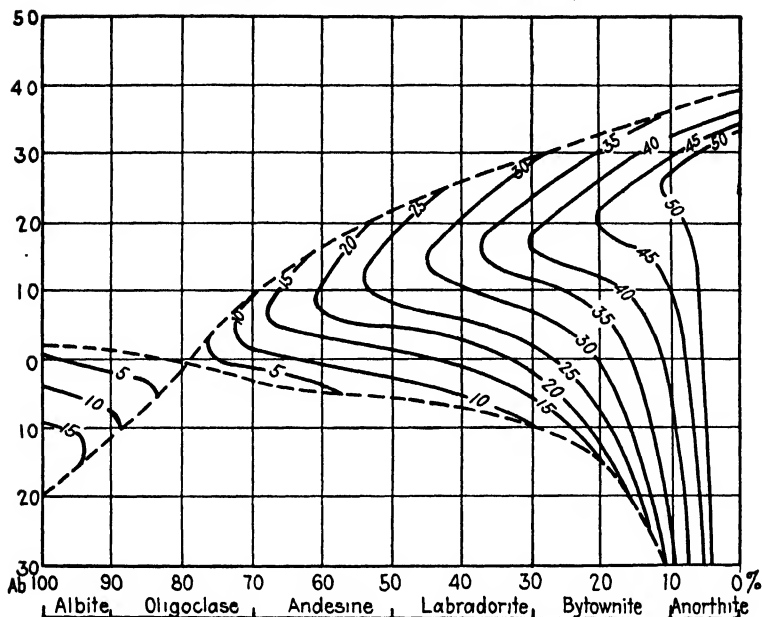


FIG. 265.—Curves of the symmetrical extinction of the bands in two parts of a Carlsbad twin of plagioclase. Ordinarily one single properly oriented crystal serves to determine the variety of plagioclase. The symmetrical extinction is determined for each half of the Carlsbad twin, and the higher value is followed in the curve of this figure until it intersects the horizontal line for the lower value. The intersection shows the composition. Only about one crystal out of ten is oriented for the lower part of the drawing, so that the curve is usually followed upward from the zero line to the intersection.

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4. *Acmite and Hornblende*.—Pyroxenes are generally distinguished from hornblende by stubby crystals, by the cleavage angle near 90° and generally by less pleochroism and colors different from hornblende. Acmite is a needle-like pyroxene with pleochroism and a green color even stronger than those of hornblende. If the cleavage angle is found, it is still a good distinction. Beyond that the interference color and the sign of elongation are characteristic. Hornblende has birefringence 0.016, elongation (+). Acmite has birefringence 0.050, elongation (—).

5. *Biotite and brown hornblende; green biotite and hornblende*.—Two cleavages indicate hornblende. If only one is seen, the hornblende usually has some faint fractures from one crack to the next, but biotite will bend rather than fracture. The interference color of biotite is the higher.

Extinction angles: $< 2^\circ$ in biotite; $< 25^\circ$ in hornblende.

$2V = 10^\circ$ in biotite; $70 - 90^\circ$ in hornblende.

6. *Cordierite* resembling quartz, orthoclase, etc., with index ranging even beyond both, has (1) yellow pleochroic haloes around zircon inclusions, (2) many sillimanite needles included and (3) structure of a penetrating trilling approaching a hexagonal form and spherulitic structure. Garnet is a characteristic associate.

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(Continued from page 469)

7. *Epidote and Augite*.—Augite has the better cleavages, and they cross at about right angles, differing in that respect from epidote. If the grains show elongation, test the position of the optic plane; it is about at right angles to the elongation in epidote but nearly parallel in most other rock minerals.

8. *Green Fibers*.—*Serpentine* in veinlets with the fibers usually transverse; often in pseudomorphs after olivine or pyroxene. *Chlorite*, in areas encroaching on various ferromagnesian minerals, forms radiating clusters, and in many cases has an ultra-blue interference color. Serpentine and chlorite have low interference colors. *Uralite* has medium and *biotite* rather high interference colors.

TABLE IV A.—ISOTROPIC MINERAL COLORS IN THIN SECTION
Names in capitals show the customary color habit of mineral

Pink	Red	Green	Blue	Brown	Yellow	Lavender or violet
GARNET Fluorite	GARNET Hauynite Chromite Perovskite	CHLORITE GREENALITE SPINEL Fluorite Garnet Hauynite Sodalite	LAZURITE NOSELITE SODALITE Fluorite Hauynite Spinel	GARNET CHROMITE SPINEL HAUYNITE GLASS Allanite Chlorite Perovskite Hisingerite Limonite	PEROVSKITE LIMONITE Garnet Hauynite Glass	Perovskite Fluorite

TABLE IV B.—ANISOTROPIC MINERAL COLORS IN THIN SECTION
Names in capitals show the customary color habit of mineral

Pink	Red	Green	Blue	Brown	Yellow	Lavender or violet
THULITE Titanite AUGITE Tourmaline Piedmontite Vesuvianite Zircon Andalusite	Biotite Rutile Bronzite Corundum Hypersthene Andalusite Titanite Hematite Staurolite Cancrinite	CHLORITE CHRYSOTILE CHLINOCHLORE HORNBLLENDE GLAUCONITE AUGITE ACMITE OTTRELITE DIALLAG ACTINOLITE CHLORITOID HYPERSTHENE Biotite Tourmaline Vesuvianite Corundum Rutile Epidote Bronzite Arfvedsonite Sillimanite Uvarovite Olivine Apatite Babingtonite Vivianite	TOURMALINE CORUNDUM GLAUCOPHANE ARFVEDSONITE SAPPHIRE LAZULITE RIEBECKITE DUMORTIERITE VIVIANITE Apatite Cordierite Axinite Chlorite Otterite Kyanite Acmite Octahedrite Vesuvianite? Chlortoid	BIOTITE TOURMALINE RUTILE ZIRCON HORNBLLENDE AXINITE ALLANITE BRONZITE BROOKITE PHLOGOPITE FAYALITE BARKEVITE Hypersthene Vesuvianite Apatite Chlorite Augite Acmite Cordierite Andalusite Monazite Staurolite Sillimanite Diallage Chondrodite Cassiterite Epidote Titanite	PHLOGOPITE CORDIERITE EPIDOTE MONAZITE STAUROLITE SULPHUR Biotite Tourmaline Rutile Chlorite Chrysotile Hypersthene Augite Hornblende Acmite Glaucophane Bronzite Actinolite Chondrodite Octahedrite Cassiterite Titanite Brookite Calcite Apatite Cancrinite Axinite Olivine Sillimanite Zircon Chloropal Humite	PIEDMONTITE Glaucophane Augite Tourmaline Rutile Axinite Vivianite

TABLE V.—MINERAL SHAPES IN THIN SECTION

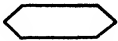





6-sided	8-sided	Rectangular	Square
			
EPIDOTE FELDSPARS HORNBLLENDE OLIVINE ZIRCON MICAS STAUROLITE Allanite	GARNET LEUCITE PYROXENES Scapolite ZIRCON	APATITE Cancrinite FELDSPARS Goethite MICAS Ottrelite Rutile TOURMALINE NEPHELITE	Apophyllite FLUORITE MAGNETITE Nephelite PYRITE Scheelite ZIRCON Spinel FELDSPAR Octahedrite
6-sided	Fibrous		Rhombic
	Needle-like	Radiating	
AMPHIBOLES APATITE Beryl Cordierite Corundum GARNET LEUCITE NEPHELITE OLIVINE QUARTZ SODALITE MICAS STAUROLITE TOURMALINE Spinel Kaolinite	Acmite APATITE Aragonite Aegirite Arfvedsonite Corundum Cancrinite CHLORITE Datolite Goethite Gypsum Rutile Sillimanite Sericite Tremolite TOURMALINE HORNBLLENDE MICAS	Chalcedony Crocidolite CHLORITE Datolite Pectolite Prehnite SERPENTINE Sillimanite TOURMALINE Uralite Wavellite ZEOLITE	AMPHIBOLES Axinite Barite CARBONATES FELDSPARS (adularia) PYROXENES TITANITE ZIRCON STAUROLITE

TABLE VI.—MINERALS THAT SHOW TWINNING

SIMPLE	POLYSYNTHETIC
(Two individuals either in contact or in partial penetration)	(Twinning planes parallel)
Allanite	Allanite
AMPHIBOLE GROUP	AMPHIBOLE GROUP
Aragonite	Aragonite
Baddeleyite (always twinned)	Baddeleyite (always twinned)
CALCITE	CALCITE
Cassiterite	Cassiterite
Clinochlore	Chloritoid (rarely observed)
Corundum	Clinochlore
Dolomite	Clinozoisite (rarely observed)
Epidote (rarely observed)	CORDIERITE
FELDSPARS	Corundum
Gypsum	DOLOMITE
Humite group	Epidote (usually not seen)
Kyanite	Epistilbite
Natrolite group	Gypsum
Olivine (rarely observed)	Humite group
Perovskite	Hydromagnesite
PYROXENE GROUP	Hydrargillite
Quartz (rarely observed)	Johnstrupite
Rutile	Kyanite
STAUROLITE	LEUCITE
Titanite	Margarite (rarely observed)
Zircon	Mica group (rarely observed)
	Mosanderite
	Natrolite group
	Ottrelite
	PLAGIOCLASE:
	Silicic tend to have narrow bands
	Basic tend to have wide bands
	Prehnite
	PYROXENE GROUP
	Quartz
	Rinkite
	Rutile
	Serendibite (always twinned)
	Titanite
	Zoisite
CROSSED, COMPOUND OR GRIDIRON (Repeated twinning after two laws)	
ANORTHOCASE	
CALCITE	
Dolomite	
LEUCITE	
MICROCLINE	
STELLATE OR CRUCIFORM (Twinning plane changes from one plane to another of the same crystal form)	
BRONZITE , etc. (also geniculated*)	
CORDIERITE	
Kyanite	
Olivine (rarely observed)	
Rutile (also geniculated*)	
STAUROLITE (also geniculated*)	
Zircon (also geniculated*)	
* Geniculated means knee shaped and when repeated tends to form a ring.	

TABLE VII A1.—OPAQUE MINERALS IN INSOLUBLE CONCENTRATES
Hardness over 5 and Specific Gravity between 2.9 and 4.0 Except as Noted

Color	Shapes (Figs. 181 and 266)	Luster	Remarks	Mineral
Strongly magnetic				
Iron-black..... Bronze..... Rusty stain on white or gray.....	Rounded octahedral grains Irregular to ragged Irregular	Metallic Metallic Metallic	G = 5.1 G = 4.5 G = 7.8 (introduced by accident)	MAGNETITE Pyrrhotite* Iron (from apparatus)*
Moderately magnetic				
Purple, brown, black..... Reddish to brown..... Brown black to black..... Black to steel gray..... Yellow brown.....	Irregular, sub-angular Rounded earthy grains Rounded octahedral, irregular, etc. Platy Stains other minerals	Metallic Earthy, metallic Metallic, submetallic Metallic, dull Earthy, submetallic	G = 5 Soluble? G = 5 Thin splinters, transparent. G = 4.5 G = 7.5 Soluble?	ILMENITE HEMATITE Chromite wolframite Limonite
Nonmagnetic				
Yellow..... Brown black..... Yellow or blue..... White, yellow, brown..... Red..... Gold yellow..... Steel white..... Black..... Gray black..... Gray black.....	Irregular, rudely grained Irregular grains, prisms Rectangular plates beveled Granular, aggregates Irregular Flattened scales, grains Grains, scales Grains, splinters Powdery, scales	Metallic Adamantine, submetallic Adamantine, submetallic Submetallic, brilliant Metallic Metallic Metallic Metallic, dull Metallic, adamantine Dull, metallic	G = 5 ± G = 7, also trans. Also trans. Pitted surface, on ilmenite H = 3. G = 9 G = 15+. H = 3 G = 19. H = 4 Soft, soluble G = 6. H = 1. G = 2.	PYRITE* (Marcasite)* Cassiterite Octahedrite LEUCOXENE copper* gold platinum pyrolusite* columbite graphite

* From alteration or addition.

The following minerals look opaque if in large grains or aggregates: **BIOTITE**, **EPIDOTE**, **KAOLINITE**, **GLAUCONITE**, **alanite**, **allimanite**, **spinelite**, **baselite**.

TABLE VII A2—TRANSPARENT ISOTROPIC MINERALS IN INSOLUBLE CONCENTRATES
Hardness More than 5 and Specific Gravity between 2.9 and 4.0 Except as Noted

Index	Color	Shapes (Figs. 181 and 266)	Remarks	Mineral
1.434 1.56 ± 6 1.61 ± 5	Violet, etc. Green, none Green	Irreg. octahedrons Irreg. shards Irreg. plates and fibers	H = 4 H = 2	Fluorite glass Chlorite
1.57 ± 1	Dull yellow to brown	Platy	Seems isotropic in plates; H = 2	Phlogopite
1.63 ± 5	Dark brown to black	Platy	Seems isotropic in plates; H = 2	BIOTITE
1.71 ± 6	Brown, black	Irreg. prisms	Some anisotropic. G = 4.1	Allanite
1.80 ± 9	Red, brown, green	Rounded, rarely dodecahedrons	Moderately magne- tic. G = 4 ±	GARNET
1.90 ± 15	Green, brown, pink	Rounded octahe- drons, pitted	G = 4.0 ±	Spinel
2.38 2.42	Yellow to brown Yellowish white	Rounded cubes Small, rounded, fac- eted grains	G = 4.1	perovskite diamond
2.47	Brown to red and black	Irregular, triang- ular?	H = 3. soluble? G = 4.0 ±, intro- duced	sphalerite

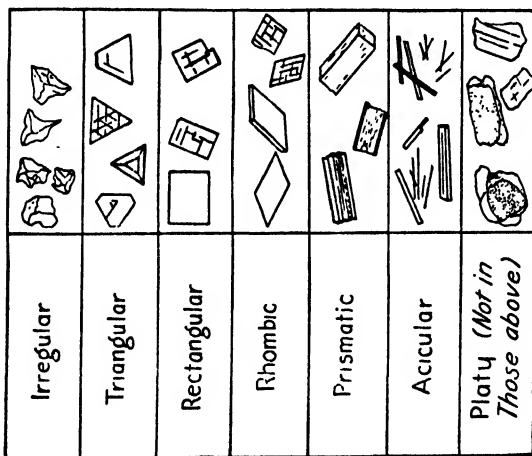


FIG. 266.—Shapes of crushed fragments.

TABLE VII A3.—ANISOTROPIC COLORLESS MINERALS IN INSOLUBLE CONCENTRATES
Hardness More than 5 and Specific Gravity between 2.9 and 4.0 Except as Noted

Beta	Birefringence	Shape		Sign	2V, degrees	Remarks	Mineral
		Sedimentary grains	Crushed fragments				
1.576	0.043	Rect. to rounded	Rect. Ir. Pl.	+	42	0° Ext. Bx. figure, Sol? H = 3.5	Anhydrite
1.58 ±	0.04 ±	Ir. to rounded plates	Ir. Pl.	-	45 ±	Coarser than assoc. concentrate.	Muscovite
1.61 ±	0.005	Rounded to Ir.	Ir.	+	0	G = 2.8, H = 2. Bx. figure	eudialite
1.62 ±	0.003	Rounded to Ir.	Ir.	-	0	+ elong.	eucolite
1.61 ±	0.05 ±	Ir. to rounded plates	Ir. Pl.	-	45 ±	Coarser than assoc. concentrate.	Muscovite
1.620	0.008	Subangular to platy	Ir. Pl.	+	65 ±	G = 2.8, H = 2. Bx. figure	TOPAZ
1.629	0.015	Pr. Pl.	-	40	Bx. figure	Wollastonite
1.634	0.003	Egg shaped	Ir. Hex.	-	0	H = 5 - G = 2.8 +. Ext. 32°	APATITE
1.637	0.012	Ir. Angular	Rect. Rh.	+	36	- elong. 0° ext.	Barite
1.638	0.011	Pr. Angular, Ir.	Pr. Ir.	+	>80	Dusty alteration. Graphitic incl.	ANDALUSITE
1.642	0.015	Pr. Ac.	Ac. Pr.	+	20	G = 4.5, H = 3	Sillimanite
1.65 ±	0.008	Pr. Ir.	Pr.	+	>45	0° ext.	Enstatite
1.656	0.05 ±	Rounded. Ir.	Ir.	+	80	Looks dirty	chondrodite
1.660	0.021	Pr. Ac.	Ac. Pr.	+	20	In contact limestones	Sillimanite
1.661	0.040	Ir. Rounded	Ir.	+	>80	0° ext.	OLIVINE
1.666	0.016	Pr. Wedges, Ac.	Pr.	+	58	Ext. 26°	spodumene
1.671	0.030	Pr. Rounded, Ir.	Pr.	+	60 ±	Ext. 45°	AUGITE*
1.685	0.032	Rh. Ir. Rounded	Ir.	+	71	In contact rocks	axinite
1.70 ±	0.007	Ir. Angular to rounded	Pr.	-	<45	Clear	EPIDOTE*
1.71 ±	0.010	Ir. Angular to rounded	Ir.	-	>70	Zoned colors	VEEVANITE
1.719	0.005	Pr. rounded	Ir.	±	0	Cl. cracks transverse, 30° ext.	KYANITE*
1.720	0.016	Pr. Subangular, Rounded	Pr. Ac.	+	>80	diaspore
1.722	0.048	Ir. Rounded	Pr.	+	>80	OLIVINE
1.750	0.050	Ir.	Ir.	-	>80	Alters to serpentine.	Corundum*
1.768	0.008 +	Pr. Rounded. Ir.	Rect. Ir.	-	0 to 10	Often blotchy G = 4.0 ±	AUGITE*
1.77 +	0.037	Slightly rounded wedges	Pr.	+	60	Ext. 45°	TITANITE*
1.907	0.134	Pr. Pyr. Egg shaped	Ir. Tetr.	+	27	Dispersion strong	ZIRCON*
1.94 ±	0.050			+	0	Zoned, G = 4.7, 0° ext.	

* Usually or commonly colored.

TABLE VII A4.—ANISOTROPIC COLORED MINERALS IN INSOLUBLE CONCENTRATES
Hardness More than 5 and Specific Gravity between 2.9 and 4.0 Except as Noted

Beta	Birefringence	Colors	Photochromism	Shape grains		Sign	2V, degrees	Remarks	Mineral
1.60 ±	01 ±	G BR g blk.	Var.	Rounded Pl.	Pl. Fibrous	±	Var.	H = 2 Bx. fig.	CHLORITE
1.60 ±	030	Pink	Str.	Jagged Pl.	Pl. Ir.	+	<20	H = 2 Bx. fig.	BIOTITE
1.61 ±	005	Pink	Var.	Rounded Pl.	Pl. Ir.	+	0	+ elong.	eudialite*
1.62 ±	003	Pink	Var.	Subangular	Ir.	+	65	+ elong.	eucolite*
1.62 ±	008	Y.	None	Rounded Pl.	Ir. Pl.	+	65		TOPAZ*
1.62 ±	011	C b y br.	Var.	Rounded Pl.	Pr.	±	Var.	H = 2	CHLORITE
1.62 ±	019	C b y br.	Pair	Pr. Ac. Fib.	Pr.	±	10° ext.	10° ext.	ACTINOLITE
1.63 ±	019	B BR v g y r blk	Str.	Pr. Ir. Rounded.	Ir.	—	0	Zoned, striated fresh	TOURMALINE
1.63 ±	004	b g.	None	Egg-shaped	Ir.	—	0	— elong 0° ext.	APATITE*
1.63 ±	011	R y v.	Var.	Pr. Ir. Angular.	Pr. Ir.	—	>80	Graphic inclusion. Dusty alteration	ANDALUSITE*
1.638	017	V B	Str.	Pr. to rounded	Pr.	—	45	Ragged	Glaucophane
1.650	017	BR y blk.	Var.	Ir. Rounded	Ir.	—	Var.	G = 4 1	Allanite
1.656	05 ±	C b y g blk.	Fair	Rounded Ir.	Ir.	+	80	Metamorphic	chondrodite
1.671	030	C b y g blk.	Slt.	Pr. Ir. Rounded.	Pr.	+	60 ±	45° ext.	AUGITE
1.685	060	BR g y.	Str.	Jagged Pl.	Pl. Ir.	+	<20	H = 2 Bx figure	BIOTITE
1.685	032	BR g y.	Fair	Rh. Ir. Rounded	Ir.	—	71	In contact rocks.	axinite
1.685	012	g y br. r.	Str.	Ragged Pr. Ir.	Pr.	—	>80	Schiller inclusions. 0° ext.	Bronzite
1.686	011	B BR v g y r blk	Str.	Pr. Ir. Rounded	Pr.	—	30 ±	Striated	dumortierite
1.690	04 ±	C b y br. blk.	Str.	Pr. Ir. frayed	Pr.	—	0	Zoned, striated fresh	TOURMALINE
1.702	031	C b y br. blk.	Str.	Ragged Pr. to Ir.	Pr.	—	60	15° + ext.	HORNBLende
1.709	004	C Y.	Fair	Unknown	Ir.	—	50-80	0° ext. Schiller inclusions	HYPERSTHENE
1.710	010	C Y.	Fair	Ir. Angular to rounded	Ir.	—	69	Metamorphic and igneous	sapphirine
1.719	005	g br.	Var.	Pr. Rounded	Ir.	—	0	Clear	Epidote
1.720	040	g br.	None	Ir. Rounded	Ir.	—	80 ±	Zoned	vesuvianite
1.720	016	B v	Slt.	Pr. Rounded.	Pr. Ac	+	>80	Alters to serpentine. Cl cracks transverse	OLIVINE*
1.721	095	y br.	Slt.	Subangular. Tetr. Pr. Pyr. Rounded.	Pr Tetr	+	0	G = 4.5. Zoned, dusty	Kyanite
1.73 ± 2	01 ±	G b blk.	Str.	Pl. warped.	Pl	+	30-60	In schists	xenotime
1.74 ±	010	Y br	Fair	Ir. Rounded	Ir.	+	>80	Many large inclusions	Chloritoid
1.74 ±	039	C Y.	Fair	Ir. Angular. Rounded	Ir. Irct	+	60-80	Clear	STAUROLITE
1.768	008	B br.	Var.	Pr. Rounded Ir	Pr	+	0 ±	Often blotchy, G = 4	EPIDOTE
1.77 ±	037	G br blk.	Slt	Ir. Rounded	Ir.	+	60 ±	45° ext.	Corundum*
1.78 ±	03 ±	BR blk.	Var.	Ir. Rounded	Ir. Pl	+	Var	G = 4 1	AUGITE and pyroxenes
1.788	031	Y g br.	Slt.	Euhedral Rounded	Ir.	+	24	G = 5 1	Allanite
1.864	050	g br.	None	Rounded wedges Pr. Pyr. Egg-shaped Pr. Pyr. Rounded Pl. Pyr	Ir. Tetr Ir. Tetr Ir. Pyr. Ir.	+	37	Alters to serpentine and Fe oxides	Monazite
1.907	134	B br y	Var.	Rounded wedges	Ir	+	0	Strong dispersion	TITANITE
1.94 ±	03 ±	BR r g v	Fair	Pr. Pyr. Egg-shaped	Ir. Tetr	+	0 ±	G = 4 6 Zoned	ZIRCON
1.997	096	y br blk.	Fair	Pl. Pyr. Rounded	Ir. Pyr.	+	0 ±	G = 7 Zoned	Cassiterite
2.554	081	y br blk.	None	Pl. Pyr	Ir.	+	<45	Striated	octahedrite
2.586	138	y br.	Slt	Pl	Ir.	+	0	Strong dispersion. Striated	Brookite
2.616	287	BR y. r. blk	Fair	Ir. Ac.	+	0	Twins Striated G = 4 2	ROUTILE

* Usually or commonly colorless.

TABLE VII B.—COMMON MINERALS OF SPECIFIC GRAVITY BELOW 2.7 OR EASY SOLUBILITY THAT CONTAMINATE MANY CONCENTRATES
Colorless Except as Noted

Gamma	Alpha	Shape		Sign	2V, degrees	Ext.	Remarks	Mineral
		Sedimentary grains	Crushed					
1.45	1.45	Ir. Rounded.	Colloidal	..	—	—	Isotropic.	Opal
1.658	1.486	Ir. Rounded.	Rh.	..	0	Sym.	Sol. Striated. Effervesces. H = 3.	CALCITE
1.681	1.500	Rh.	Rh.	—	0	Sym.	Sol. Effervesces. H = 3	DOLOMITE
1.685	1.530	Ir. Fossils.	Pr. Radial.	—	18	0°	Sol. Effervesces. H = 3	aragonite
1.526	1.519	Pl. Rounded.	Pr. Pl. Ir.	—	0-70	6°	See chart, dusty	ORTHOCLASE
1.529	1.522	Pl. Rounded. Ir.	Pr. Pl. Ir.	—	83	15°	See chart, gridiron twins	MICROCLINE
1.56 + 3	1.55 ± 3	Ir. Rect. Rounded.	Pr. Pl. Ir.	±	80 ±	Var.	See chart, albite twins	PLAGIOCLASE
1.530	1.521	Fib. Rad. Rounded. Crystals. Rounded.	Pl. Rh.	+	58	25°	H = 2. Sol. Striated	Gypsum
1.537	1.530	Angular. Ir.	Ir. Ac.	±	Var.	Colloform. Fractured	CHALCEDONY
1.59 ± 3	Dusty, Oolitic	±	Earthy. H = 1 to 4.	Bauxite
1.544	1.535	Ir. Rounded.	Ir.	..	60 ± 20	Inclusions sillimanite, etc.	Cordierite
1.553	1.544	Ir. Rounded. Crystals.	Ir. Angular	+	0 ±	{ Clear to slightly dusty { Ext. commonly wavy. }	QUARTZ
1.57 ± 1	1.56 ± 1	Ir.	Ac. Ir.	+	Var.	0°	H = 4. Green, etc. Inclusions Fe oxide	Serpentine*
1.600	1.560	Rounded. Pl.	Pl.	—	40 ±	0°	{ Bx. figure. H = 2 { Larger than associates }	MUSCOVITE
1.567	1.561	Hex. Pl. Ir. Dust.	Pl. Ir. Dust	—	60	0°	H = 1. Earthy	KAOLINITE
1.628	1.610	Round. Ir. Fossils.	Ac.	—	20	H = 2 +. Green fib. gran- ules.	Glauconite
1.61 ± 5	1.60 ± 5	Ir. Pl.	Pl. Ac.	±	Var.	0°	H = 2. Green	Chlorite
1.87 +	1.63 +	Rh. to rounded.	Rh.	—	0	Sym.	{ Some Rad. Sol. Efferves- ces. H = 3 }	SIDERITE
Opaque	Dust	Flakes?	Black. G = 2. H = 1	Graphite

* Secondary after olivine and pyroxene.

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